A PALAEOENVIRONMENTAL ANALYSIS OF THE ARCHAEAN MOODIES GROUP, BARBERTON MOUNTAIN LAND, SOUTH AFRICA.

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ABSTRACT

The Moodies Group, approximately 3 300 m.y. in age, is the oldest relatively unmetamorphosed quartzitic assemblage of sediments presently known. This succession consists of a wide variety of sedimentary facies which are interpreted palaeoenvironmentally in terms of interacting alluvial, estuarine-deltaic, beach-offshore and tidal flat depositional systems. A remarkable similarity exists between Holocene physical sedimentary processes and those operative during the Archaean.

Proximal outwash fan sediments consist of conglomerates with minor sandstones. Matrix-supported conglomerates display good correlation between bed thickness and maximum clast-size, suggesting a mass flow mode of origin. Clast-supported conglomerates, in contrast, exhibit no such correlation, and represent traction deposits on longitudinal bars. Associated plane-bedded sandstones formed on bar surfaces at shallow-water depths under upper flow regime conditions. Mid-alluvial plain sandstones are structured by northward-directed large- and medium-scale trough cross-beds and contain occasional scattered pebbles. These deposits represent braided bar accumulations during falling stages of episodic floods. Cycles are capped either by plane-bedded sandstones or by ripple-drift cross-laminated sandstones and siltstones. These formed respectively on bar surfaces under upper flow regime conditions during a rapid lowering of water-level, and at lower energies with appreciable quantities of sediment deposited from suspension. Associated in-phase ripple-drift cross-laminated siltstones with shale drape-laminae represent overbank or levee accumulations. The most distal deposits consist of upward-fining, channel-fill sequences enclosed within horizontally-laminated and ripple-cross-laminated siltstones and shales of probable flood-plain origin.

Three deltaic facies are recognized. The first consists of upward-fining cycles of conglomerates and sandstones with thin shale partings, enclosed within rhythmically interlayered argillaceous sandstones and shales. These sediments

are interpreted as tidal channels which meandered across estuarine tidal flats, and constitute the tidal deltaic plain. The second facies comprises lensoid orthoquartzitic sandstone bodies, with ebb-oriented planar and herringbone cross-bedding passing upwards into plane-bedded sandstones with desiccated shale partings. These are interpreted as subtidal and intertidal sand shoals. Associated poorly-sorted felspathic sandstones, devoid of primary sedimentary structures but containing water-escape pipes and other penecontemporaneous deformational structures throughout their development, are the third facies. The evidence of extensive dewatering is indicative of rapid sedimentation, and, in association with the two overlying facies, suggests a delta front depositional environment. The above criteria indicate that the delta as a whole was subjected to high tidal energies, and had the form of a macrotidal estuary fronted by elongate sand shoals.

Barrier island and shallow shelf sediments are arranged in upward-coarsening progradational sequences. Low energy offshore deposits consist of banded iron formations overlain by shales with jasper layers and lenses. Siltstones become more abundant upwards as chemical sediments disappear in the nearshore depositional environment. Interbedded fine-grained sandstones are interpreted as storm-surge deposits. Shoreface accumulations consist of trough cross-bedded sandstones deposited by predominantly offshore-directed rip currents. These are overlain by upward-fining orthoquartzitic tidal inlet cycles which contain basal scattered pebbles passing upwards into unimodal planar, and herringbone cross-beds which decrease in set thickness upwards and contain shale flasers on foresets. Thin siltstone layers cap some of the tidal inlet cycles. Overlying plane-bedded sandstones with low-angle discordances are less pure than the shoreface and tidal inlet deposits. Coupled with the presence of abundant rippled surfaces, thin shale partings and longshore directed planar intrasets a barrier spit, as opposed to a foreshore depositional environment, is suggested.

Back-barrier tidal deposits are comprised of progradational upward-fining cycles of herringbone cross-bedded and flaser bedded sandstones and w vy and lenticular bedded sandstones and shales with various types of ripples, desiccation cracks and rill marks. Associated large-scale unimodal cross-bedding, medium-scale bimodal-bipolar cross-bedding, and plane-bedding probably formed in a complex of flood tidal delta subenvironments. Intercalated washover fan deposits consist of thick sequences of plane-bedded sandstones.

A depositional model, to relate the different sedimentary environments in space, has been suggested. Braided alluvial plain and deltaic sedimentation represent the constructive elements of the model. Marine reworking of riverborne sediments le to the development of barrier islands and back-barrier tidal deposits lateral to the deltaic plain. The presence of spit and abundant rip current deposits in the barrier island sequences, coupled with the absence of extensive ebb tidal delta accumulations, is indicative of a microtidal coastline. Local embayments of the coastline, however, resulted in macrotidal conditions under which tidal channel and tidal sand shoal deposits developed in a non-barred estuary. Low energy and storm-surge sedimentation in the nearshore and suspension and chemical sedimentation in the offshore complete the depositional model.

Vertical sequences in the Moodies Group were determined by source area tectonics, and can be accounted for by prograding or transgressing the steady state model. Tidal flat deposits only developed under transgressive conditions. Barrier island sequences, in contrast, required a regressive shoreline for their preservation. During active progradation, back-barrier tidal flat sedimentation did not occur as alluvial plain sediments often built directly onto shelf accumulations in the form of fan deltas. Marginal destructive swash bars developed on fan deltas at the cessation of fluvial influx.

While providing no new clues as to the original composition of the earth's crust, the nature of the Moodies sediments indicates the existence of widespread exposed granitic terrains by 3 300 m.y. In addition, the abundance of marginal marine orthoquartzitic sandstones in the Moodies Group suggests a relatively stable crust on which extensive reworking of sediments occurred. The proposed micro- and only local macrotidal ranges, however, imply a fairly narrow shelf during deposition of the Moodies Group. The conformably underlying Fig Tree sediments, which are considered to be a deep-water facies equivalent of the Moodies Group, are further indicative of a narrow shelf. The above criteria have been used to suggest that the Moodies and Fig Tree groups accumulated along an ancient continental margin which probably had the geometry of a marginal basin. These generalities appear to have characterised Archaean sedimentation in South Africa and contrast, in many respects, with conditions which prevailed during the Proterozoic. Sedimentary successions on the Kaapvaal Craton, such as the Pongola Supergroup and Pretoria Group, are not only devoid of deep-water sediments but also contain widespread evidence of broad, shallow-shelf macrotidal sedimentation. Marginal marine facies are lacking in the Archaean of Canada, which contains alluvial and turbidite sediments deposited in localized trough-like basins. The two Archaean styles of sedimentation may reflect responses to different stages of continental rifting. Further work on other Archaean sedimentary sequences is, however, necessary to test this hypothesis.

CONTENTS

INTRODUCTION.

i) Objectives and Approach.

An appreciation of how physical sedimentary processes and depositional environments in the Archaean compare with those in later geological times and in the Holocene can provide valuable information pertaining to the evolution of the Earth's crust. The presence of mature beach and tidal sediments, for instance, can be taken to imply a stable crust on which extensive reworking of sediments occurred. Conversely, a predominance of immature deep-water greywacke assemblages would suggest a relatively unstable setting. These contrasting styles are typified by sedimentation in intracratonic basins and in basins adjacent to orogenic belts. If both sediment-types occur as facies equivalents it is probable that sedimentation took place along a continental margin. Detailed studies on tidal sediments, and more specifically on tidal ranges as a function of shelf width (Redfield, 1958), may provide additional evidence as to the nature of depositional basins. Tidal sediments, furthermore, are significant with respect to the origin of the Earth-Moon system.

A complete understanding of the depositional history of a sedimentary basin requires a knowledge of the interaction of major depositional environments not only along strike but also down the palaeoslope. Fortunately, Archaean sediments are usually tightly folded into parallel synclines and anticlines and therefore particularly amenable to sedimentary facies analysis. This involves the following approach :

a) The recognition of different sedimentary facies, and their definition in terms of grain/pebble-size, texture (Powers, 1953; Folk, 1968), composition (modified after Dott, 1964; chert is here considered to be a stable component in quartz arenites and not a rock fragment), sedimentary structures and vertical sequences;

b) A palaeoenvironmental interpretation of the different facies with reference to Holocene processes;

c) An evaluation of the interrelationships of the facies within specific depositional environments with reference to Holocene depositional models;

d) An analysis of the interrelationships of the different environments to yield a single depositional model; and,

e) An account of vertical stratigraphic relationships in terms of the depositional model.

ii) Previous Work on Archaean Sediments.

The most common Archaean sedimentary facies consists of graded beds of greywacke-shale, often with sole marks and slump features. These sediments are interpreted as turbidites, and are considered to reflect deposition in a fairly deep-water environment. Associated conglomerates and massive sandstones are interpreted as 'fluxoturbidites' and 'proximal turbidites'(Walker, 1970). This assemblage of sediments thus contains material resedimented from shallow- into deep-waters (turbidites and fluxoturbidites) and normal background deep-water deposits (shales, cherts and banded iron formations) and is referred to as the 'Resedimented Association' (Turner and Walker, 1973). Sedimentary sequences representative of this association have been reported from Archaean greenstone belts in Canada (Donaldson and Jackson, 1965; Walker and Pettijohn, 1971; Henderson, 1972; Pettijohn, 1972 and Turner and Walker, 1973), the U.S.A. (Ojakangas, 1972 and Granath, 1975), Australia (Glikson, 1971; Dunbar and McCall, 1971; Lipple, 1974; Fitton, et al., 1975; Hallberg et al., 1976; and D.I. Groves, personal communication, 1977), Rhodesia (Bliss and Stidolph, 1969) and South Africa (Anhaeusser et al, 1968; Anhaeusser, 1973 and Reimer, 1975).

The second Archaean sedimentary facies contains, amongst other lithologies and sedimentary structures, conglomerates, impure and orthoquartzitic sandstones with large-, medium- and small-scale cross-beds, and shales. This facies, generally considered to be a shallow-water association, is less common than the first and has only been reported from a few greenstone belts. Prior to 1973, Reimer (1967), Anhaeusser et al. (1968), Anhaeusser (1969), and Glikson (1971) had made brief reference to shallow-water sedimentary structures, including mud-cracks, in the Archaean of South Africa and Australia respectively, while Donaldson and Jackson (1965) had suggested that extensive shallow-water accumulations, subsequently removed by erosion, may have existed in northwest Ontario. Orthoquartzitic pebbles from the same area (J.A. Donaldson and R.W. Ojakangas, personal communication, 1974) may provide indirect evidence of shallow-water depositional conditions with extensive reworking. Detailed sedimentological studies on sediments from this facies were rare until Turner and Walker (1973) provided conclusive evidence of shallow-water sedimentation in the Archaean when they interpreted a conglomeratic and arkosic sandstone occurrence in Canada as a subaerial alluvial fan deposit. The term 'Continental Association' was proposed for sediments of this type with large- and medium-scale cross-beds considered to be the best evidence for shallow water deposition. The most recent studies on sediments belonging to the second facies are from the Yilgarn Block of Western Australia where Donaldson and Platt (1975) interpreted an upward-coarsening assemblage as a prograding deltaic, braided river and alluvial fan sequence, and where Marston and Travis (1976) recognised alluvial fan deposits of very local derivation. Similar sediments exist in the eastern (Lipple, 1974) and western (Fitton, et al., 1975) portions of the Pilbara Block, as well as in Rhodesia (Bliss and Stidolph, 1969; Bickle et al., 1975), but no palaeoenvironmental analyses have been undertaken.

The Moodies Group in the Barberton Mountain Land, South Africa (Anhaeusser et al., 1968; Anhaeusser, 1973) is a well-preserved sequence of sediments of the shallow-water type. Studies on this unit, as developed in the southeastern Transvaal and northern Swaziland, have been exclusively of a stratigraphic nature (Hall, 1918; Visser, 1956; Jones, 1963; Bell, 1967; Reimer, 1967; Tomlinson, 1967; Jones, 1969 and Anhaeusser, 1976), apart from the documentation of shallow-water sedimentary structures (Reimer, 1967; Anhaeusser, 1969; Anhaeusser et al., 1973 and Anhaeusser, 1976). The present investigation focuses initially on the sedimentological history of the Moodies Group before extending the findings into a general analysis of Archaean sedimentation and the related crustal evolutionary implications.

STRATIGRAPHIC-FRAMEWORK AND AGE OF THE MOODIES GROUP.

The Moodies Group is the uppermost of three subdivisions of the Swaziland Supergroup (Fig. 1). The basal Onverwacht Group consists predominantly of volcanics with minor sediments (see Viljoen and Viljoen, 1969; Anhaeusser, 1973; amongst others) and is conformably overlain by immature clastic and chemical sediments of the Fig Tree Group (Reimer, 1975). In the southern parts of the Barberton Mountain Land, most notably in Swaziland, the Moodies Group rests directly on Onverwacht volcanics (Jones, 1969) and in the central parts,unconformably on Fig Tree sediments. It was based on these field relationships that earlier workers such as Hall (1918) and Visser (1956) considered the Moodies Group as a separate 'System' from the underlying Onverwacht and Fig Tree 'Series' which were grouped together as the Swaziland System. From the Makonjwa Range northwards (Fig. 1), as was established by Tomlinson (1967) and Anhaeusser (1969), a gradational contact exists between the Fig Tree and Moodies groups. This led Viljoen and Viljoen (1969) to establish the Onverwacht, Fig Tree and Moodies as groups within the Swaziland Supergroup.

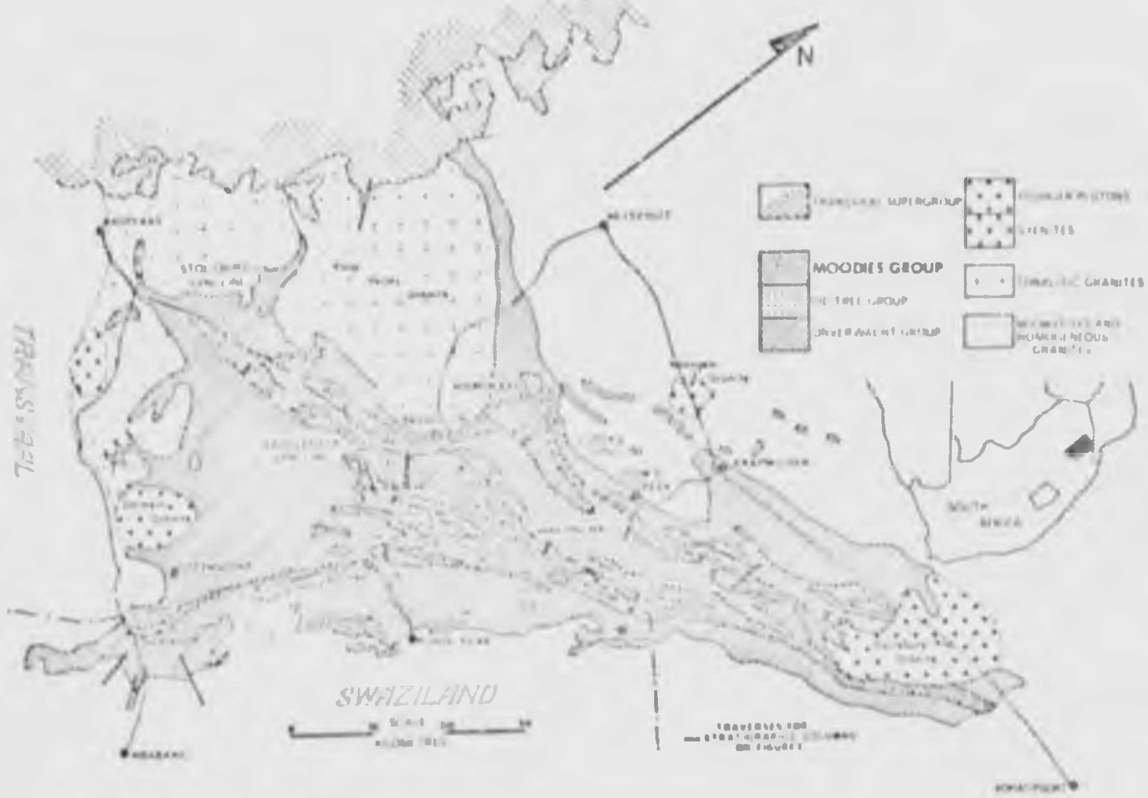

Fig. 1. Geological map of the Barberton Mountain Land with inset map of South Africa (modified after Truswell, 1970; after Anhaeusser et al., 1968).

No direct age determinations have been carried out on the Moodies Group rocks, but these sediments can be dated relative to upper Onverwacht felsic volcanics, (with an age of 3 360 ± 100 m.y., Van Niekerk and Burger, 1969) at less than 3 400 m.y. A minimum age for the Moodies sediments is more difficult to determine. The 'intrusive' nature of the Kaap Valley Granite, which has been dated at 3 360 ± 100 m.y. (Oosthuyzen, 1970), is now disputed, and other more reliable evidence must be sought. As the Swaziland Supergroup is strongly deformed, the age of non-foliated or post-tectonic granites will provide a minimum age for the Moodies Group. Granites of this type include the intrusive Lochiel or Homogeneous Hood Granite, dated at 3 070 ± 60 m.y. (Allsopp et al., 1962) and the intrusive Salisburykop Pluton, which has an age of 3 060 ± 30 m.y. (Oosthuyzen, op. cit.). The oldest non-foliated granite is the Dalmein Pluton which is dated at 3 290 ± 80 m.y. (Oosthuyzen, op. cit.). This suggests that the Moodies Group is at least 3 300 m.y. in age which, in view of the conformable nature of the three groups within the Swaz nd Supergroup, is not incompatible with the age of the upper Onverwacht. The Moodies Group is thus older than any of the sedimentary basins on the Kaapvaal Craton. The oldest of these is the Pongola Supergroup which rests on granites of the Lochiel-type, dated at 3 000 ± 200 m.y. (H.L. Allsopp, personal communication, 1975).

The Moodies Group attains its greatest thickness in the Eureka Syncline (Fig. 1), where it is subdivided into five stratigraphic units, termed MD1 through MD5 (Fig. 2). A rominent amygdaloidal lava zone at the base of unit MD4 allows correlation between the Eureka, Saddleback and Stolzburg Synclines (Fig. 1). In the latter two synclines the Moodies Group is represented by stratigraphic units MD1 through MD4. South of the Saddleback Syncline, no correlatable stratigraphic units, apart from a basal conglomerate, can be recognized (Fig. 2).

In the southern parts of the Barberton Mountain Land, notably along the Swaziland border and northwards to the Saddleback Syncline (Fig. 1), conglomerates and impure sanastones are the predominant lithologies (Fig. 2). Intercalated orthoquartzitic sandstones and a penecontemporaneously deformed sandstone unit also occur in the Saddleback Syncline. Orthoquartzitic sandstones, in association with thick accumulations of shales and siltstones, are present in the Eureka and Stolzburg Synclines, where quartzose sandstones are also abundantly developed.

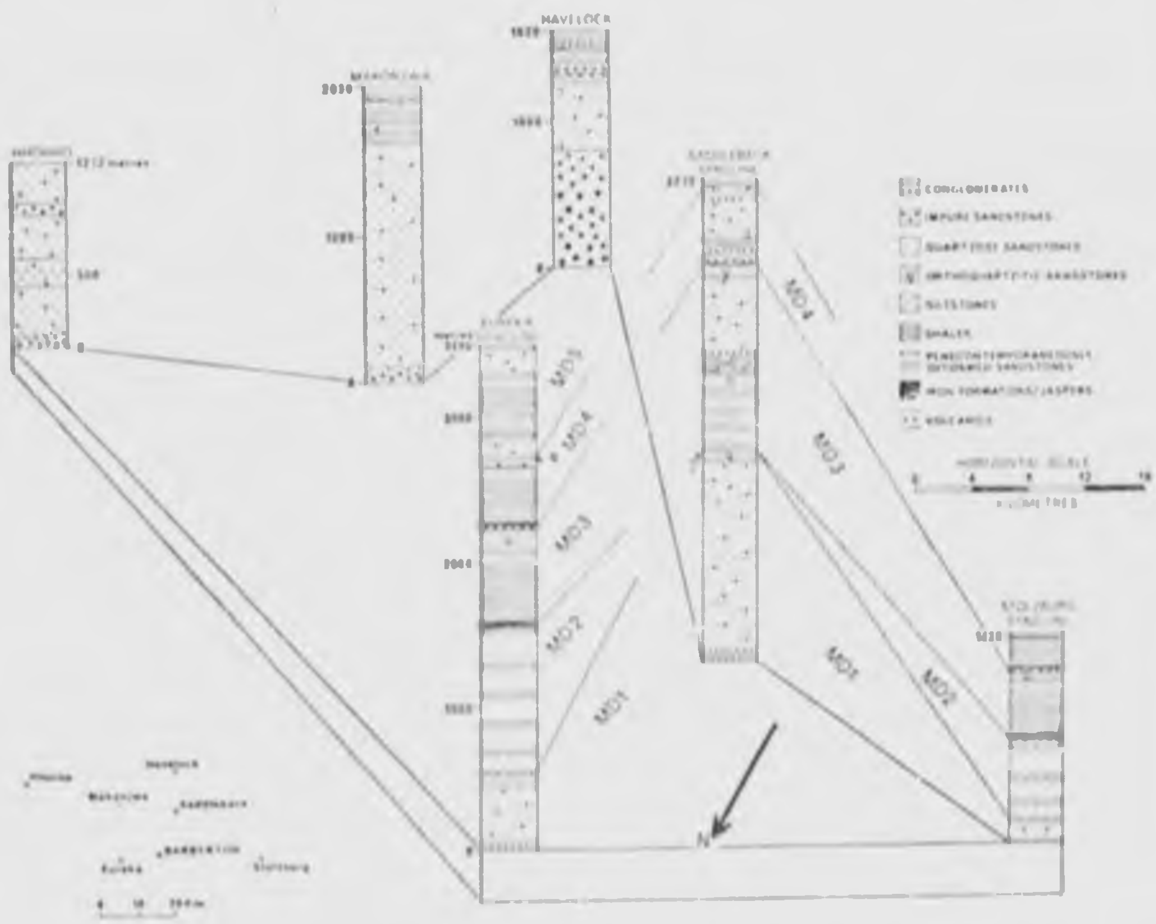

Fig. 2. Representative stratigraphic columns of the Moodies Group. Inset shows ground positions of column localities relative to Barberton.

RECOGNITION OF MARINE SEDIMENTS IN THE PRECAMBRIAN.

The most reliable indicators of marine depositional environments are metazoans, either as shelly or trace fossils. Shelly fossils do not occur in Precambrian sediments and the oldest documented trace fossils are now dated at approximately 1 000 m.y. (Clemmey, 1976). In rocks older than 1 000 m.y. other criteria for marine deposition must be sought.

i) The Origin of Orthoquartzites.

The possible significance of orthoquartzites as indicators of marine depositional environments has been reviewed by Horne and Ferm (1976) who emphasise that each occurrence should be treated independently, since clean quartzose sandstones have three possible modes of formation. Firstly, orthoquartzites may form diagenetically. As a result of intense weathering, leaching of less stable minerals may occur and result in the residual accumulation of quartz. Pure quartz arenites formed by this process tend to be restricted in their distribution, and are most commonly developed in soil zones. Orthoquartzitic deposits may also be derived by erosion of pure quartz source areas. Under these circumstances, both continental and marine sandstones would be orthoquartzitic. Widespread orthoquartzites most commonly accumulate in depositional environments where intense reworking of sediment results in winnowing of finer-grained materials and the breakdown of minerals less stable than quartz. This most commonly occurs in marginal marine settings as a result of reworking by tides (Swett et al., 1971; Balazs and Klein, 1972) and waves (Hunt, 1887). Swett et al. (op. cit.) have shown that sand grains may be moved for up to 36 km/year by tidal currents, either in subtidal or intertidal environments, or both. Similar distances of transport can probably be invoked for sand particles in offshore shoals. These figures contrast with distances of transport and resultant abrasion in fluvial environments, where Kuenen (1959) showed that abrasion of subangular quartz grains, less than 2 mm in diameter, is less than 1%/1000 km of transport. Orthoquartzitic marine sandstones will thus interfinger with relatively impure but time-equivalent continental sandstones.

ii) Sediment Dispersal Patterns.

Detailed mapping of palaeocurrents can aid considerably in the recognition of Precambrian marine sediments. In contrast to the dominantly unidirectional currents which operate in alluvial settings, marginal marine environments are usually characterised by multidirectional dispersal processes. Directional structures are generated by tidal, wave- and wind-driven currents and may be oriented down, up, parallel or diagonal to the slope of the sedimentary basin (Klein, 1967a). Multidirectional dispersal patterns in the geological record can thus be used to infer marginal marine or marine environments. Furthermore, if the general attitude of the palaeoslope can be determined through palaeocurrent mapping of fluvial facies it may be possible to relate associated, more varied, dispersal patterns to currents which operated in specific marginal marine environments.

iii) Vertical Sedimentary Sequences.

Idealized vertical sequences of lithologies and sedimentary structures are well established from the Holocene for braided fluvial (Boothroyd and Ashley, 1975), meandering fluvial (Bernard et al., 1970), barrier island (Bernard et al., 1962), deltaic (Coleman and Wright, 1975), and tidal flat (Klein, 1972) depositional environments. The recognition of similar sequences in Precambrian successions provides a further means of recognizing ancient depositional environments, especially if considered in conjunction with the previously discussed lithological and palaeocurrent criteria.

DEPOSITIONAL FACIES ASSEMBLAGE 'A'

i) Introduction.

This is the dominant facies assemblage in the Moodies Group and is characterised by abundant conglomerates intercalated within thick sequences of impure sandstones, and subordinate shales. In the southern parts of the Barberton Mountain Land, notably along the Swaziland border and northwards to the Makonjwa Range (Fig. 1), this is the only facies assemblage present (see Fig. 2). In the Saddleback Syncline stratigraphic units MD1, the upper part of MD3, and MD4 consist of facies assemblage 'A'. Further north representatives of this assemblage are less common and comprise only MD1, the upper part of MD4, and MD5 in the Eureka Syncline, and MD1 and the upper part of MD4 in the Stolzburg Syncline. Three facies are distinguishable in this assemblage.

ii) Description of Facies.

A1. Conglomerate Facies.

Occurrences of this facies are indicated on Fig. 2, from which it is apparent that the thickest development of conglomerates occurs in the south, and specifically around Havelock. The conglomerates are very coarse, with an average mean pebble-diameter of -5, 1Ø (3,3 cm) (Table 1), but often contain boulders greater than 25 cm in diameter. Pebbles and boulders are generally spherical and rounded to well-rounded except in the Mhohho area (Fig. 1) where a thin conglomerate with angular to subangular clasts occurs at the base of the succession. Sorting (dispersion) is commonly poor (> 1,0) and rarely moderate (,70 - 1,0) (Folk, 1968, see Table 1). White, black and banded cherts and jasper are the predomin c clast-types in the conglomerates (Table 2). Acid volcanic clasts, often silicified, are fairly abundant, but mafic varieties are rare. Granitic clasts, often greater than 25 cm in diameter, are confined to the basal conglomerate and occur most commonly in the Eureka Syncline.

TABLE 1.

STATISTICAL PARAMETERS FOR MOODIES CONGLOMERATES

Sample No.	E1	E2	E3	E4	S1	S2	S3	S4
Mean (Ø)	-4,88	-5,25	-5,07	-4,56	-5,41	-5,62	-5,40	-4,65
Variance (Ø)	1,03	1,41	1,82	1,17	1,37	2,40	1,58	,55
Std. Dev. (Dispersion) (Ø)	1,01	1,19	1,35	1,08	1,17	1,55	1,26	,74
Max. Clast Size (cm)	12	30	50	24	33	45	38	16
Number of Measurements	119	129	108	108	111	127	110	120

E1 - E4 : Basal Conglomerate, Havelock Area

S1 and S2 : Middle MD3 Conglomerate, Saddleback Syncline

S3 : Lower MD4 Conglomerate, Saddleback Syncline

S4 : Upper MD4 Conglomerate, Saddleback Syncline

(See Fig. 2 for stratigraphic positions)

Statistical parameters are calculated for grouped data with frequency count by class. All clasts >1 cm in diameter were measured at each locality.

Equations :

$$\mu\,(Mean) = \frac{\sum_{i=1}^{k} f_i x_i}{N}$$

$$\sigma^2 (Variance) = \frac{\sum_{i=1}^{k} f_i x_i^2 - \dfrac{\left(\sum_{i=1}^{k} f_i x_i\right)^2}{N}}{N-1}$$

(Table 1. continued)

where : x_i = Midpoint of Class Interval

f_i = Frequency in Each Class

k = Number of Classes

N = Number of Observations

(after Edwards, 1964, p. 41 and 59)

TABLE 2.

CLAST TYPES IN MOODIES CONGLOMERATES

Sample Number	E1	E4	S1	S3	S4
White Chert	10,5	1,9	25,2	45,5	8,1
Black Chert	26,8	48,2	30,0	1,8	58,5
Banded Chert	39,0	25,0	13,5	18,8	8,9
Vein Quartz	4,9	1,9	,9	1,8	4,1
Acid Volcanic	9,8	6,5	9,0	16,1	8,2
Mafic Volcanic	-	,9	3,6	8,1	1,6
Ironstone	3,2	2,8	17,1	6,5	-
Jasper	4,1	13,0	-	-	2,4
Brecciated Chert	1,6	-	,9	1,0	-

Figures in Percentages

Samples as in Table 1.

Two conglomerate-types occur in this facies; one is matrix-supported (Fig. 3) and the other clast-supported (Fig. 4). Matrix-supported types are most common lower down in the sedimentary succession especially in the southern parts of the outcrop belt. The basal conglomerate in the Havelock area, for instance, is predominantly matrix-supported with only occasional thin intercalated clast-supported units. Clast-supported conglomerates are more

Fig. 3. Matrix-supported conglomerate: Havelock area (scale in 1 and 5 cm divisions).

Fig. 4. Imbricated, clast-supported conglomerate: Unit MD4; Saddleback Syncline.

abundant higher up in the stratigraphic succession, notably in unit MD4 in the Saddleback and Stolzburg synclines (Fig.2). Contacts between sedimentary units, as defined by thin intercalated sandstones, are regular and non-erosional in the matrix-supported variety, while the clast-supported type frequently displays irregular basal contacts which may have a relief of up to 10 m. Imbrication is not apparent in the matrix-supported conglomerates and is rarely seen in the clast-supported types (Fig. 4). The matrix-supported conglomerates display no size-grading of clasts, but an upward-fining is often noticeable in the clast-supported varieties. The relationship between bed thickness and maximum clast-size for four different conglomerates is illustrated in Fig. 5. Correlation is best for conglomerates which are clearly matrix-supported (Havelock and MD2 Saddleback) and very poor for the lower MD4 clast-supported variety (Fig. 5). A fair correlation exists for the upper MD4 occurrence which contains both matrix- and clast-supported conglomerates.

The matrix of the conglomerates consists of medium- to coarse-grained, poorly- to very poorly-sorted lithicarenites and sublitharenites. The intercalated pebble-free sandstones are similar in composition to the matrix. These are most commonly plane-bedded (Fig. 6 and 7.) and rarely greater than 1 m in thickness. They persist laterally for tens of metres where associated with the matrix-supported conglomerates, but tend to pinch out along strike where intercalated within the clast-supported variety (Fig. 7). Trough cross-bedded sandstones vertically separate some individual conglomerate units of the clast-supported type, and often grade laterally into these conglomerates.

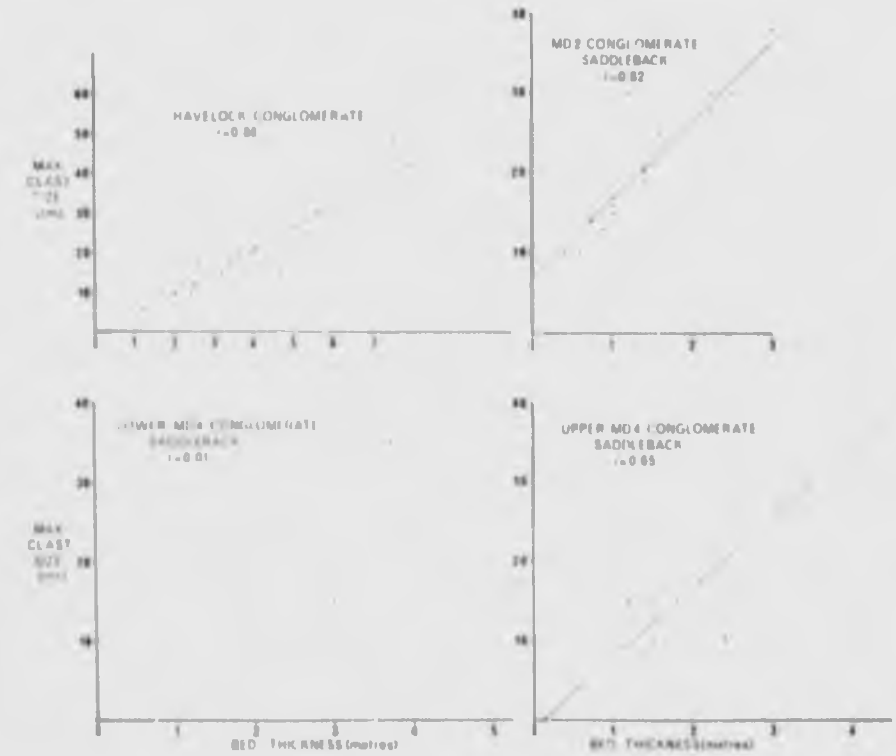

Fig. 5. Graphic plot of bed thickness against maximum clast-size for four different conglomerates (r = correlation coefficient).

Fig. 6. Plane-bedded sandstones associated with clast-supported conglomerate: Unit MD4; Saddleback Syncline.

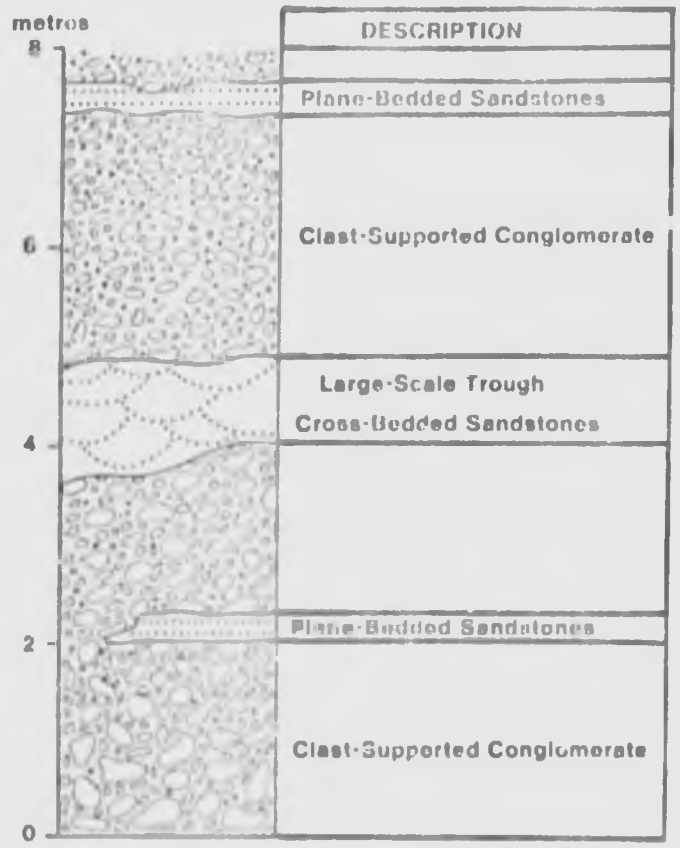

Fig. 7. Measured stratigraphic section through clast-supported conglomerates and associated sandstones: upper conglomerate; Unit MD4; Saddleback Syncline.

The two end-member conglomerate-types are best typified by the Havelock and lower MD4 conglomerates, respectively. Frequently, however, the matrix- and clast-supported conglomerates display a rapid alternation in pairs varying between 4 and 10 metres in thickness. This is particularly noticeable in the upper MD4 conglomerate of the Saddleback area (Fig. 2).

A2. <u>Trough Cross-Bedded Sandstone Facies</u>.

Sediments belonging to this facies are grouped together on Fig. 2 as 'impure sandstones'. The sandstones vary in composition from sublitharenites and litharenites to lithic greywackes, are fine- to coarse-grained and poorly-to very-poorly sorted. Cumulative frequency curves (Fig. 8) and statistical parameters (Table 3) for a number of samples from this facies indicate their highly immature character. Scattered pebbles, predominantly of chert, are common but siltstone and shale layers constitute a very small proportion of this facies.

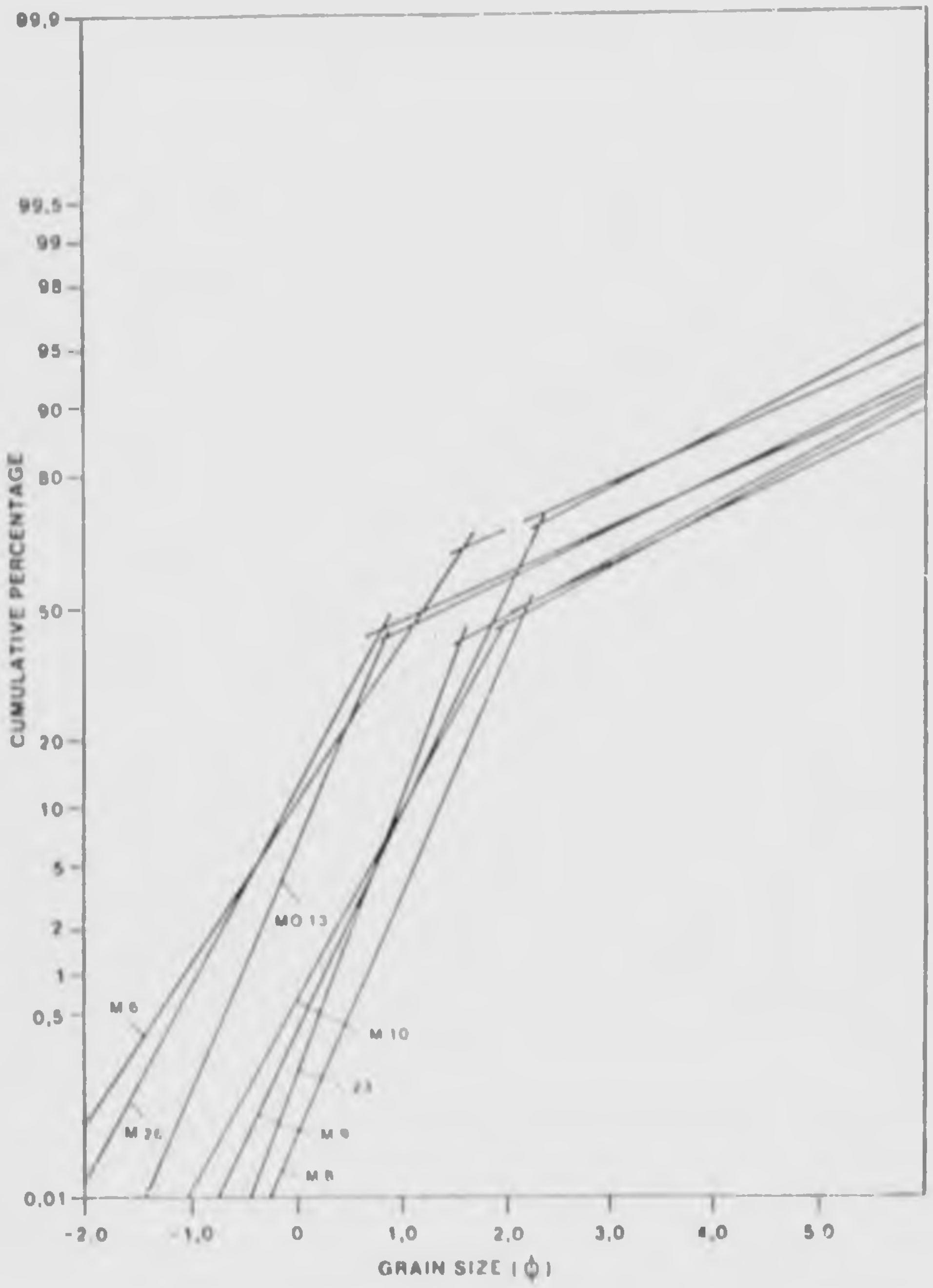

Fig. 8. Grain-size cumulative frequency curves for 'impure sandstones'. See Table 3 for sample localities. Curves drawn for converted sieve equivalent percentiles.

TABLE 3.

STATISTICAL PARAMETERS FOR MOODIES IMPURE SANDSTONES

Sample No.	M8	M6	MO13	23	M26	M9	M10
Mean (Ø)	3,35	2,03	2,43	3,03	3,18	2,36	3,18
Variance (Ø)	3,24	2,86	4,33	3,76	3,72	1,54	3,76
Std. Dev. (Ø)	1,80	1,69	2,08	1,93	2,21	1,24	1,93
Number of Measurements	200	200	200	200	200	200	200

M8 : Unit MD1, Saddleback Syncline

M6 : Unit MD3, Saddleback Syncline

MO13 : Unit MD4, Saddleback Syncline

23 : Matrix to Lower Conglomerate in Unit MD4, Saddleback Syncline

M26 : Makonjwa Area

M9 : Makonjwa Area

M10 : Makonjwa Area

(See Fig. 2 for stratigraphic positions)

Longest diameters of 200 grains were measured in thin section for each sample. Statistical parameters were calculated using the equations of Inman (1952) after converting cumulative frequency thin section percentiles to sieve equivalents (conversion equations after Harrell and Eriksson, in preparation).

See Appendix I for Statistical Parameter and Thin Section - Sieve Conversion Equations.

Examples of vertical sequences of sedimentary structures developed in this facies are illustrated in Fig. 9. Depositional cycles vary from 5 to 30 m in thickness, and often commence with a scattered pebble layer. Medium- and large-scale trough and occasional planar cross-beds (Figs. 10, 11 and 12) are the predominant sedimentary structures in overlying sandstones. Troughs are up to 6 m in width and vary from 20 cm to 2,5 m in thickness. Markedly erosional contacts, which may be occupied by shale-partings, separate the trough cross-bed sets. Shale-partings and flakes occur along some cross-bed foresets. Depositional cycles are capped by a variety of sedimentary structures including small-scale cross-bedding or ripple-cross-lamination, plane-bedding with or without primary current lineations, ripple-drift cross-lamination and desiccated shale-partings, shale drape-laminae (Fig. 13) and convolute lamination in siltstone (Fig. 14). The thick sequences of 'impure sandstones' (Fig. 2) consist of stacked, overlapping and often scour-based partial or complete cycles of the types illustrated in Fig. 9. Where not fully developed, the upper parts of cycles are missing, and thick uninterrupted intervals of trough cross-bedded sandstones are developed.

Palaeocurrent measurements on medium-scale trough cross-beds were made at a number of stations on different stratigraphic units. The cumulative results of these measurements for each stratigraphic unit are shown in Fig. 15 * (See Figs. 1 and 2 for localities). A consistent northerly transport direction is indicated throughout the deposition of this facies.

*FOOTNOTE 1 : See Appendix II for field measurement and stereonet rotation procedures.

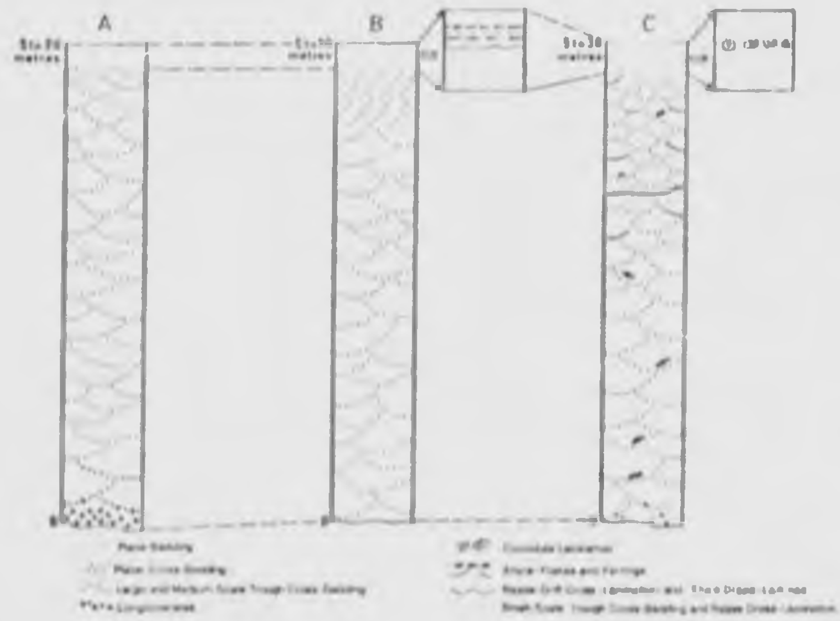

Fig. 9. Measured vertical sequences of lithologies and sedimentary structures in the trough cross-bedded sandstone facies.

A3. <u>Impure Sandstone - Siltstone - Shale Facies.</u>

Sediments of this facies are somewhat restricted in their distribution, and are illustrated in Fig. 2 as siltstones and shales in the Makonjwa and Havelock areas, part of unit MD1 and much of MD5 in the Eureka Syncline, limited portions of unit MD4 in the Saddleback Syncline and part of unit MD1 in the Stolzburg Syncline. Thin conglomerate-sandstone sequences, which always constitute less than 10 percent of measured stratigraphic intervals, are often present within the predominant alternating siltstones and shales.

The coarse-grained sediments of this facies are arranged in 2 to 6 m thick upward-fining sequences (Fig. 16). Thin scattered-pebble and shale-flake conglomerates pass

Fig. 10. Large-scale trough cross-beds: Makonjwa area.

Fig. 11. Large-scale trough cross-beds: Unit MD1; Saddleback Syncline.

Fig. 12. Planar cross-beds; Unit MD1; Saddleback Syncline.

Fig. 13. In-phase ripple-drift cross-lamination and shale drape-laminae overlain by base of succeeding depositional cycle: Unit MD3; Saddleback Syncline.

Fig. 14. Convolute lamination and ripple-drift cross-lamination: Unit MD5; Eureka Syncline.

Fig. 15. Cross-bed vector rose diagrams for the trough cross-bedded sandstone facies.

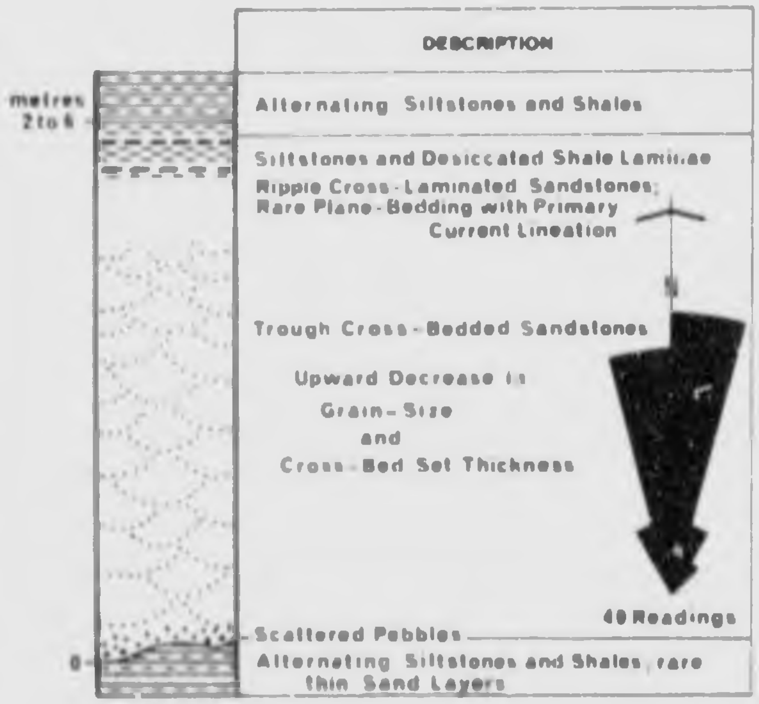

Fig. 16. Measured stratigraphic section through impure sandstone-siltstone-shale facies: Makonjwa area.

Fig. 17. Finely-bedded alternating siltstones and shales: Unit MD4; Saddleback Syncline. Note the graded, upward-fining arrangement of siltstone-shale pairs especially above the hammer handle.

upwards into lithic greywackes structured by northerly-directed small- and medium-scale trough cross-beds. These sandstones decrease in grain-size upwards from coarse at the base to medium- and fine-grained at the top of the cycles. An upward decrease in cross-bed set thickness, from 30 to less than 10 cm, also occurs. The upper fine-grained sandstones occasionally display plane-bedding with primary current lineations, and are in turn overlain by siltstones with thin desiccated shale-partings. Single upward-fining cycles, which generally extend laterally for tens of metres, are enclosed within horizontally-laminated and less commonly ripple-cross-laminated siltstones and horizontally-laminated shales (Fig. 17). The latter are commonly arranged in graded upward-fining units between 1 and 5 cm in thickness.

iii) Palaeoenvironmental Interpretation.

a) General Environment.

A number of features of this facies assemblage can be used in a general palaeoenvironmental interpretation. The most important of these are the immature character of the sediments, the strongly unimodal palaeoccurent patterns, and the widespread evidence of desiccation. These criteria indicate an environment of limited reworking which was influenced by unidirectional dispersal currents and subjected periodically to subaerial exposure, and are most compatible with an alluvial depositional setting. The shape of the cumulative frequency curves, specifically the high suspended population (Fig. 8; Visher, 1969) as well as the vertical sequences of lithologies and sedimentary structures developed in this facies assemblage (Figs. 7, 9 and 16) further support an alluvial depositional environment (see for example Klein, 1972; Boothroyd and Ashley, 1975).

The three facies of this assemblage are now analysed in the context of the broadly defined alluvial plain depositional environment.

b) Conglomerate Facies (A1)

The two readily distinguishable conglomerate-types developed in this facies must have originated in different sedimentary subenvironments and as a result of contrasting depositional processes within the general alluvial setting.

The coarseness of the matrix-supported conglomerates, their poor sorting, lack of grading and stratification, and the general absence of associated cross-bedding as well as the consistent lateral thickness of individual beds, are all suggestive of a debris-flow mode of origin (Blissenbach, 1954; Bull, 1972; Walker, 1975). The arenaceous matrix to these conglomerates, however, argues against a true debris flow origin. Similar sand-supported conglomerates have been ascribed by Miall (1970) to 'debris flood' processes which are essentially sand debris flow phenomena. The less specific term 'mass flow' is preferred for these matrix-supported conglomerates which contain no direct evidence for a viscous, matrix-strength support mechanism during their transport (Middleton and Hampton, 1973). Individual matrix-supported conglomerate beds are thought to be related to separate mass flows, which were probably generated during successive floods. The bed thickness to maximum clast-size graphic plots (Fig. 5) provide further information as to the process involved in the deposition of the matrix-supported conglomerates. Bluck (1967) showed that conglomerates deposited by torrential floods exhibit a good correlation between those two variables. This is due to the fact that more intense flows can transport greater quantities of sediment and larger clasts which are deposited almost instantaneously as single bedding units. The excellent correlation between bed thickness and maximum clast-size in the matrix-supported conglomerates from this facies indicates that individual bedding units can be related to separate mass flows. The minor interbedded sandstones were probably formed during waning stages of individual floods.

Although coarse-grained and poorly-sorted, the clast-supported conglomerates of this facies are frequently well-stratified and often display internal grading and a weak imbrication. In addition, abundant channeling occurs and the intercalated sandstones are frequently cross-bedded (Fig. 18). These features suggest that traction was the important process involved in the deposition of these alluvial conglomerates and associated sandstones. The poor correlation between clast-size and bed thickness in the lower MD4 conglomerate (Fig. 5) further supports this hypothesis.

The upper reaches of modern braided alluvial plains are frequently structured by longitudinal bars composed of pebbles, cobbles and boulders (see, for example, Doeglas, 1962; Boothroyd, 1972; Gustavson, 1974; Boothroyd and Ashley, 1975). These bars are most commonly initiated by deposition of the coarse bedload fraction of a stream as lags in the middle of the channel (Leopold and Wolman, 1957; Rust, 1972). The height of the bars is determined by fluid and sediment discharge (Hein and Walker, 1977). If both remain high after deposition of the lag the bar will grow downstream faster than it aggrades vertically. A rapid decrease in both fluid and sediment discharge conversely results in vertical aggradation of the bar. Holocene gravel bars are generally capped by sands which, depending on the flow regime, may be ripple-cross-laminated, small-scale cross-bedded or plane-bedded (Rust, op cit.). Channels adjacent to the gravel bars frequently contain lower flow regime megaripples which continue to migrate at low flow stages after gravel movement has ceased (Williams and Rust, 1969; Boothroyd and Ashley, op cit.). The clast-supported conglomerates and associated sandstones of this facies (Fig. 18) can be explained in terms of these Holocene processes. Variations in fluid and sediment discharge probably determined the thickness of conglomerate beds (Fig. 5). Open gravel frameworks developed during relatively high discharge with the matrix deposited during waning-flow. The predominance of plane-bedding in the intercalated sandstones implies short-lived upper flow regime conditions associated with rapid lowering of water level (Boothroyd and Ashley, op. cit.). Intercalated trough cross-bedded sandstones are thought to represent low-water channel deposits over which gravel bars migrated as they shifted laterally (Doeglas, 1962).

The alternating matrix- and clast-supported conglomerates can be interpreted as due to fractional, possible sheetflood (Bull, 1972) reworking of earlier mass flow deposits. The moderate correlation between bed thickness and maximum clast-size in the upper MD4 conglomerate of the Saddleback Syncline may be a reflection of these two depositional processes, and contrasts with the excellent and poor correlations of the end member Havelock and lower MD4 occurrences, respectively.

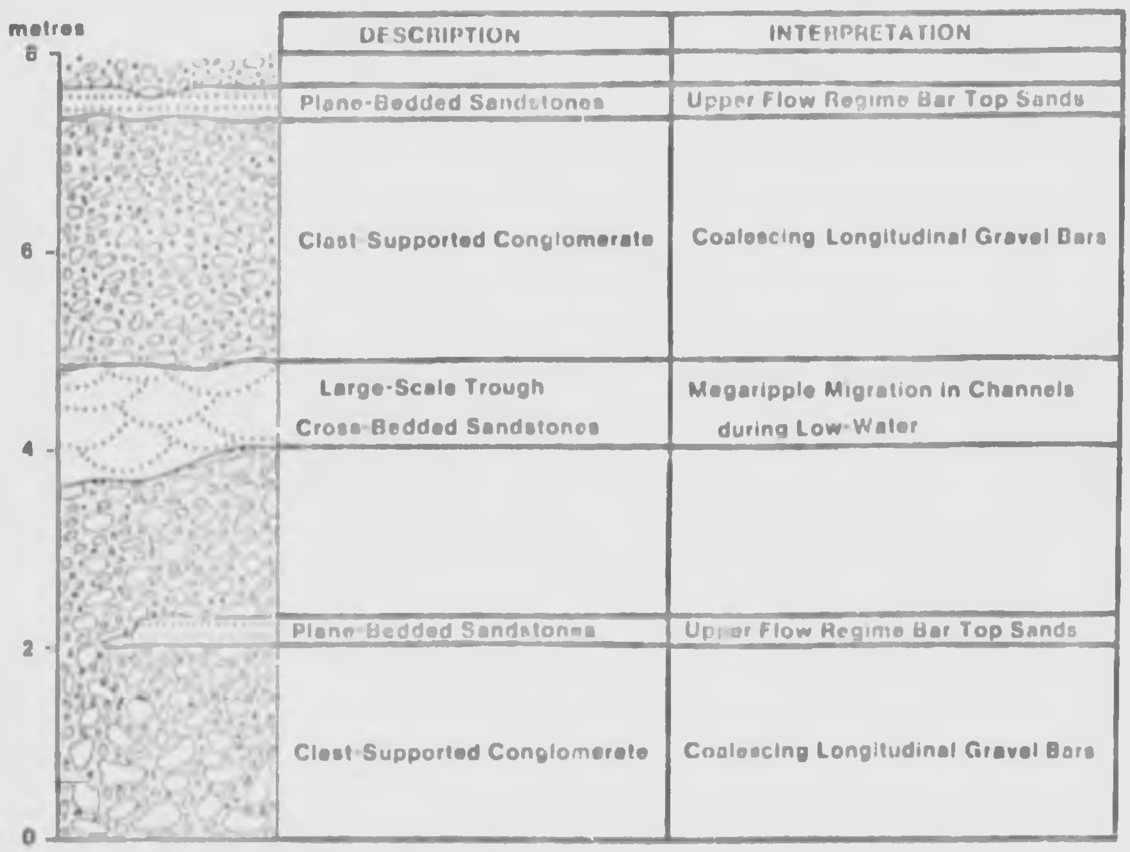

Fig. 18. Palaeoenvironmental interpretation for clast-supported conglomerates and associated sandstones.

c) <u>Trough Cross-Bedded Sandstone Facies (A2)</u>

Upward-fining cycles similar to those illustrated in Fig. 19 form by channel filling and lateral migration in modern braided stream environments (Allen, 1965; Coleman, 1969; Boothroyd and Ashley, 1975). The braided channels of the Brahmaputra River are characterised by innumerable sandbars and mid-channel islands, which develop when the stream load exceeds its carrying capacity. Sandbars are diamond-shaped in plan view, with long axes parallel to the flow, and have their longer downstream faces covered with ripples and larger bedforms. Vertical accumulations of cross-bedded sandstones, up to 18 m in thickness, are deposited in relatively short periods of time and are attributed to deposition by a migrating sand bar during a single flood cycle. As a result of rapid lateral migration of the thalweg, a large percentage of these sand units are preserved throughout the length of the river.

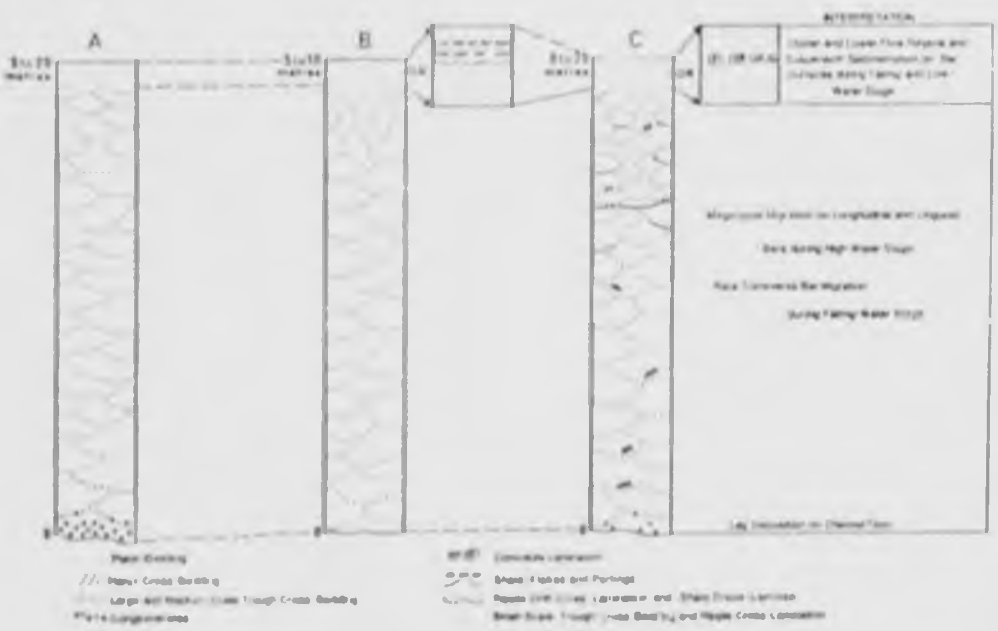

Fig. 19. Palaeoenvironmental interpretation for trough cross-bedded sandstone facies.

Similar depositional processes, including considerable vertical accretion, can be invoked for the thick sequences of cross-bedded sandstones in this facies (Fig. 19). Migration of megaripples across the surfaces of linguoid or longitudinal bars, resulting in the formation of trough cross-beds, was the dominant depositional process. Downstream accretion of transverse bars probably accounts for the occasional planar cross-beds (Fig. 19B). The abundance of trough cross-bedding suggests that the streams had a low-sinuosity (Moody-Stuart, 1966), a hypothesis which is supported by the strongly unimodal palaeocurrent patterns (Fig. 15). Low sinuosity streams also characteristically develop channel-fills by vertical accretion, rather than lateral migration (Moody-Stuart, op. cit.). Basal conglomerates are interpreted as channel lags which formed during highest discharge and over which sandbars accreted as the stream velocity decreased.

Migration of large-scale bedforms occurs primarily during high-water stage. With decreasing discharge and resulting drop in water-level a variety of hydraulic regimes are generated on bar surfaces. These are reflected in the diverse assemblage of sedimentary structures which cap depositional cycles. Plane-bedded sandstones (Fig. 19B) formed under upper flow regime conditions as a response to rapid lower [illegible] he water-level (Boothroyd and Ashley, 1975). Wit [illegible] adually decreasing flow velocities, high-amplitude gave way to low-amplitude bedforms, leading to the development of small-scale cross-beds or ripple-cross-lamination at the top of depositional cycles (Fig. 19A). Capping ripple-drift cross-lamination (Fig. 19B and C) formed when abundant sand and silt were deposited from suspension. The gradation from type B to in-phase ripple-drift cross-lamination (Fig. 19C) reflects an [illegible]reasing suspended load to bedload ratio (Jopling and Walker, 1968) under waning flow velocities (Gustavson, et al., 1975). Associated convolute lamination (Figs. 19C and 14) developed during dewatering of the inherently saturated ripple-drift cross-laminated sands and silts. In-phase ripple-drift cross-lamination and associated shale drape-laminae are similar to overbank levee accumulations in modern braided alluvial plains (Boothroyd and Ashley, op. cit.) and represent the only such deposits in this facies. Overbank sedimentation was confined to abandoned reaches of braided alluvial plains.

Although vertical accretion was an important process involved in the development of this facies, the persistence of the trough cross-bedded sandstones both along strike and down the palaeoslope indicates that downstream and lateral migration were likewise significant. Sandbars in the Brahmaputra River migrate downstream for up to 1700 m during a single flood, at rates of between 90 and 120 m per day. The same river has been found to migrate laterally for distances of over 700 m in short periods of time (Coleman, 1969). Even greater lateral migration rates are available for the Kosi River for which figures of up to 30 km per year have been obtained (Fahnestock, 1963). Braided floodplains are thus highly active depositional regimes consisting of active and abandoned channels. Through constant shifting of stream courses, thick vertical accumulations of sediments are developed. Erosion during the lateral shifting of channels is responsible for the frequent removal of the bar surface deposits (Fig. 19) resulting in the formation of thick sequences of overlapping, scour-based trough cross-bedded sandstone lenses of the type present in this facies.

d) <u>Impure Sandstone-Siltstone-Shale Facies. (A3)</u>

The character of the sediments in this facies indicates deposition in two contrasting fluvial subenvironments (see review by Allen, 1965). In terms of their textures, sedimentary structures and vertical sequences (Fig. 20), the conglomerates and sandstones are analogous to channel sediments of modern streams (Harms and Fahnestock, 1965; Sarkar and Basumallick, 1968). The enclosing finer-grained sediments closely resemble contemporary overbank floodplain deposits (McKee et al., 1967).

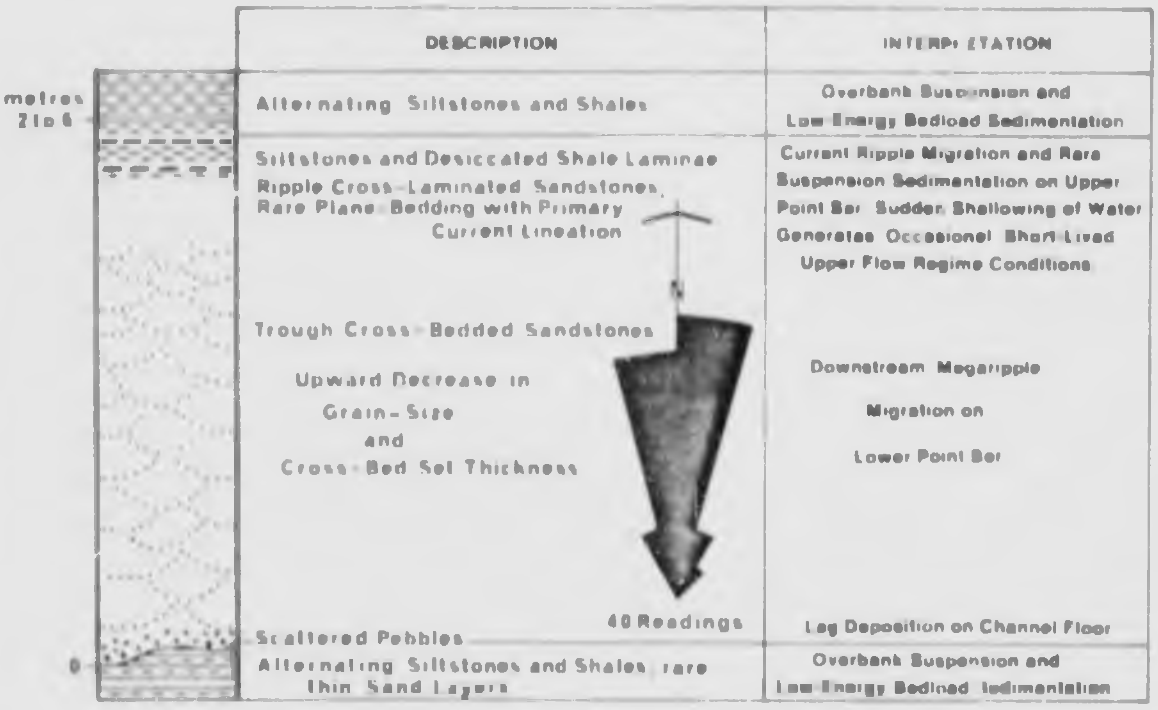

Fig. 20. Inferred depositional environments of impure sandstone-siltstone-shale facies.

Their persistence along strike, and limited thickness, suggest that lateral rather than vertical accretion was the most important process responsible for the deposition of the conglomerates and sandstones of this facies. Meandering rivers most commonly undergo this type of lateral migration as a result of erosion of the outer and deposition on the inner bank (Visher, 1972). Vertical changes in depositional structures and grain-sizes within the coarse-member can be attributed to the spatial variation of hydraulic parameters within river channels (Allen, 1970). Basal scattered-pebble conglomerates represent channel lag deposits across which lower point bar coarse- and medium-grained trough cross-bedded sandstones accreted (see Fig. 20). A continuing decrease in flow velocity at shallower water depths resulted in the development of finer-grained sandstones, structured by small-scale trough cross-beds and ripple-cross-lamination, on the upper point bar. Occasional primary current lineations on plane-bedded fine-grained sandstones indicate short-lived upper flow regime conditions on the upper point bars as a result of a decrease in water depth.

The fine-grained member of this facies developed under waning bedload and suspension sedimentation processes and, as indicated by the absence of desiccation cracks, were probably maintained in a continually saturated state. The predominance of ripple-cross-lamination and plane-bedding in the siltstones and fine-grained sandstones, to the exclusion of large-scale sedimentary structures, is in accord with experimental findings. Willis et al. (1972) have shown that, for grain-sizes of less than 0,10 mm, increasing flow velocities result in ripples giving way directly to upper flat beds. Wolman and Leopold (1957) in turn found that high velocities are common during overbank flows.

In many respects the depositional model proposed for this facies corresponds to the high-sinuosity model of Moody-Stuart (1966). Although extensive large-scale planar cross-beds are not developed, lateral point bar migration was still an important process involved in the formation of the upward-fining cycles. Furthermore, the thickness of the coarse-grained units indicate that channel-depths varied between 2 and 6 metres (Schumm, 1972; Leeder, 1973).

e) Discussion.

Having interpreted each facies in terms of local processes it is now necessary to develop a single depositional model for this assemblage of fluvial sediments. In general terms, alluvial plains are characterised by a downstream decrease in grain-size, especially for sediments which constitute bars (Smith, 1974; Boothroyd and Ashley, 1975), and a downstream variation in bar morphology. Channels in the upper reaches of alluvial plains generally contain longitudinal or linguoid bars, while those in lower alluvial plains are more commonly structured by transverse or point bars (Smith, 1970; Boothroyd and Ashley, op. cit.,). Braided channels develop on steeper proximal slopes and meandering channels on low gradient, more distal parts of alluvial plains (Leopold and Wolman, 1957). A sympathetic change from low- to high-sinuosity streams occurs away from the source area.

Based on the above, it can be implied that the conglomeratic facies of this assemblage was deposited on the upper, more southerly, and the impure sandstone-siltstone-shale facies on the lower, more northerly reaches of the alluvial plain. Thick sequences of trough cross-bedded sandstones accumulated in mid-alluvial plain environments.

The exact nature of the upper alluvial plain, on which conglomerates were deposited by both mass flow and tractional processes, requires further clarification. Debris flows most characteristically occur on dry region alluvial fans and result in the formation of muddy, matrix-supported conglomerates (Bull, 1972). Sandy, matrix-supported conglomerates of the type present in this facies, are atypical of alluvial fans, and may have developed instead on the upper parts of humid region outwash fans during floods. Inter-relationships between the clast-supported conglomerates and associated sandstones are similar to those for gravel bar and channel bar sediments on the upper reaches of the Scott and Yana fans in Alaska (Boothroyd and Ashley, op. cit.) Both conglomerate-types thus probably accumulated on upper outwash fans but with the matrix-supported conglomerates deposited proximal to the clast-supported types (Hooke, 1967). The frequent interlayering of the two conglomerate-types, indicates that mass flow and tractional processes alternated through time on the upper outwash fans.

DEPOSITIONAL FACIES ASSEMBLAGE 'B'

i) Introduction.

This facies assemblage is limited to the lower half of unit MD3 in the Saddleback Syncline (Fig. 2). Lithologic characterisation is not possible, as conglomerates, impure sandstones, orthoquartzites, siltstones and shales occur within the assemblage. Three distinct facies in sharp lithologic contact can be recognized. Two of these are continuous throughout the Saddleback Syncline; the third (medium- to coarse-grained sandstone facies) has a more restricted distribution and, where present, separates the other two. The facies are described stratigraphically from top to bottom.

ii) Description of Facies.

B1. Conglomerate-Sandstone Facies.

This facies, which occurs as a laterally persistent unit up to 300 m in thickness, consists of a series of stacked lensoid bodies of conglomerate and sandstone enclosed in finer-grained sediments. Individual lenses are composed of a number of 1-3 m thick upward-fining cycles (Fig. 21). A basal conglomerate is composed of pebbles of chert, vein quartz, quartzite and lava, from 1 to 8 cm in diameter, and angular shale-flakes, in a coarse subarkosic matrix. The pebbles vary considerably in their degree of rounding and sorting. Overlying coarse-grained subarkoses contain abundant unidirectional, and less common bidirectional planar cross-bedding (Fig. 22), commonly with shale-drapes or small pebbles on the foreset surfaces. Flow separation is indicated by regressive ripples (Fig. 23; Boersma, 1967) in a few of the planar cross-bed units which also contain reactivation surfaces. The upper parts of these cycles consist predominantly of plane-bedded, fine- to medium-grained quartz wackes with thin shale-partings or flasers (Fig. 24). These shale-partings are invariably desiccated (Fig. 25) with angular and rounded (Fig. 26) flakes reworked into overlying sediments. Small-scale planar and herringbone cross-bedding (Fig. 21, see also Fig. 22) are also developed in this unit. Thin units of fine-grained ripple-drift cross-laminated sandstones and siltstones (Fig. 27) often occupy the upper parts of these cycles and include types A and B which rarely become in-phase (terminology of Jopling and Walker, 1968). The coarser-grained lenses are enclosed in rhythmically interlayered dark mudstones and highly argillaceous coarse- to fine-grained quartz wackes. Desiccation is common, and the wackes invariably contain angular shale-flakes.

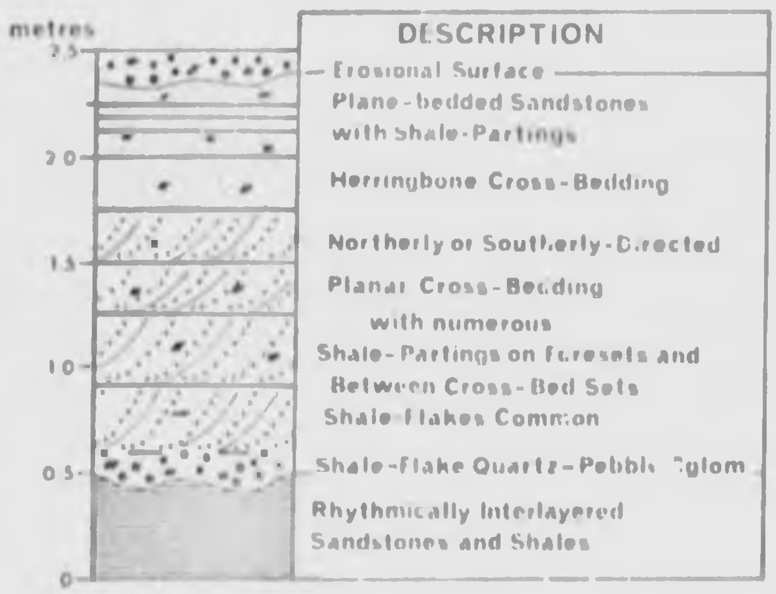

Fig. 21. Idealized stratigraphic column through conglomerate-sandstone facies: Unit MD3; Saddleback Syncline.

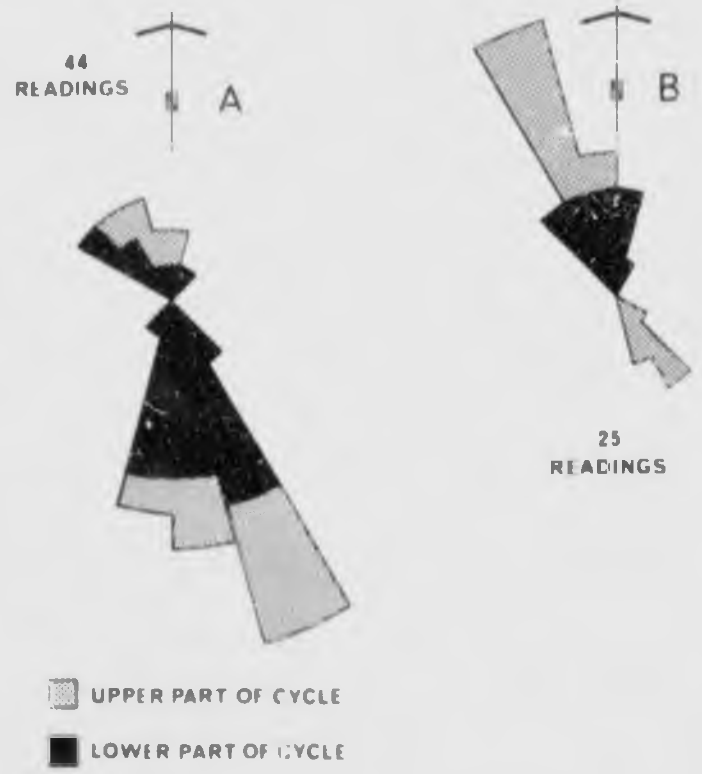

Fig. 22. Cross-bed vector rose diagrams for conglomerate-sandstone facies: Unit MD3; Saddleback Syncline.

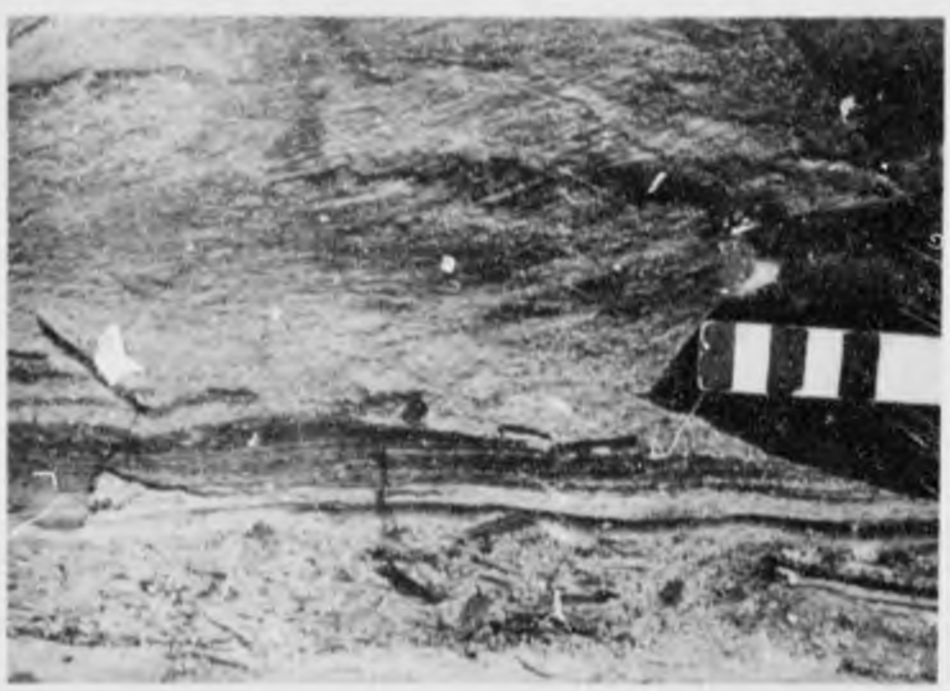

Fig. 23. Planar cross-bed set with regressive ripples to left of scale and shale-partings on foresets: Unit MD3; Saddleback Syncline (scale in 1 cm divisions).

Fig. 24. Plane-bedded sandstones with shale-partings: Unit MD3; Saddleback Syncline.

Fig. 25. Desiccated shale-partings: Unit MD3; Saddleback Syncline.

Fig. 26. Rounded shale-flakes: Unit MD3; Saddleback Syncline (note pen top for scale).

Fig. 27. Type A ripple-drift cross-lamination: Unit MD3; Saddleback Syncline (scale in 1 and 5 cm divisions).

B2. Medium- to Coarse-Grained Sandstone Facies.

Medium- to coarse-grained quartz arenites are developed through an 80 m thickness (Fig. 28), and along a strike length of 4 km, in the western part of the Saddleback Syncline (Fig. 1). They are characterised by high chert and polycrystalline quartz contents, which together may constitute up to 25 per cent of the rock. Grains vary from well-rounded to rounded to sub-rounded, and the sediments are moderately- to well-sorted. Occasional pebble streaks occupy the bases of erosional channels in the sandstones. Scattered pebbles are also common, varying from sub-angular to well-rounded, and averaging 5 cm in diameter. They consist mainly of black and white chert with subordinate vein quartz, quartzite and silicified lava. Rip-up shale-flakes are also present.

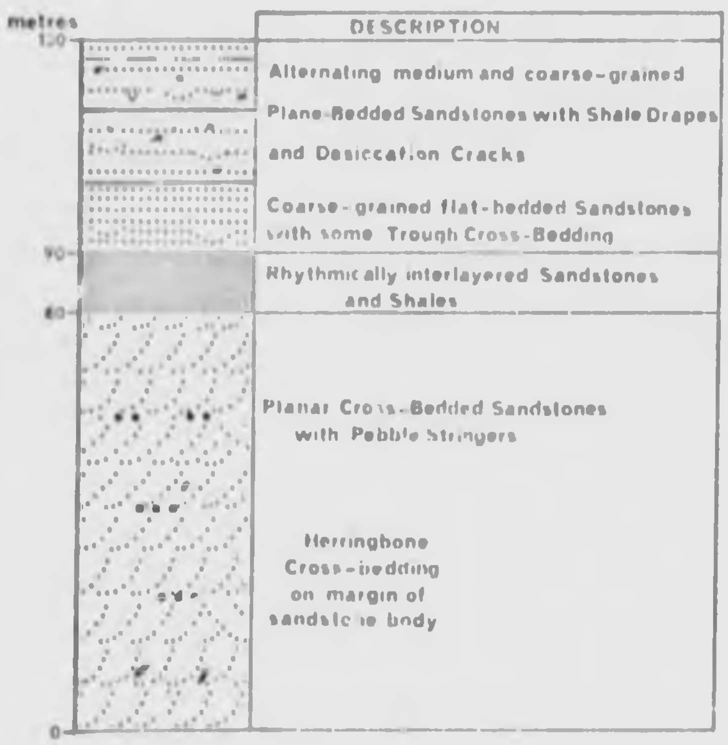

Fig. 28. Measured stratigraphic section through medium- to coarse-grained sandstone facies: Unit MD3; Saddleback Syncline.

Fig. 29. Stacked unimodal planar cross-bed sets: Unit MD3; Saddleback Syncline (note portion of hammer head for scale).

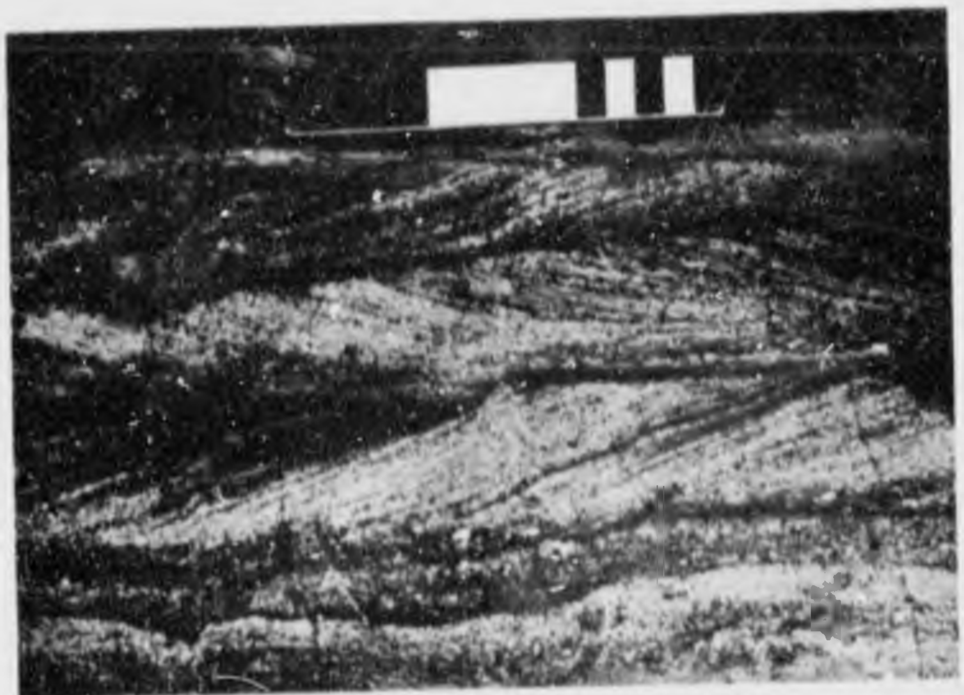

Fig. 30. Herringbone cross-bedding: Unit MD3; Saddleback Syncline (scale in 1 and 5 cm divisions).

In its lower 80 m this facies consists of laterally persistent planar cross-bedded units with pebble streaks (Fig. 28). Individual cross-bed units average 15 cm, and are up to 30 cm, in thickness. The cross-bedding is generally unimodal (Fig. 29) towards the northeast quadrant, with less common southwesterly oriented foresets (Fig. 31A). On the exposed eastern margin of the sandstone body, however, herringbone cross-bedding (Fig. 30) predominates, resulting in distinct bimodal-bipolar dispersal patterns (Fig. 31B). Other sedimentary structures include reactivation surfaces (Klein, 1970a) and linear ripples. Unusual occurrences of ripple-drift cross-lamination in medium- to coarse-grained sandstones [illegible] o charact[illegible] this facies. Types A and B occur to the exclusi[illegible] -phase variety. Deformed foreset laminae are r[illegible] emporaneous deformation features.

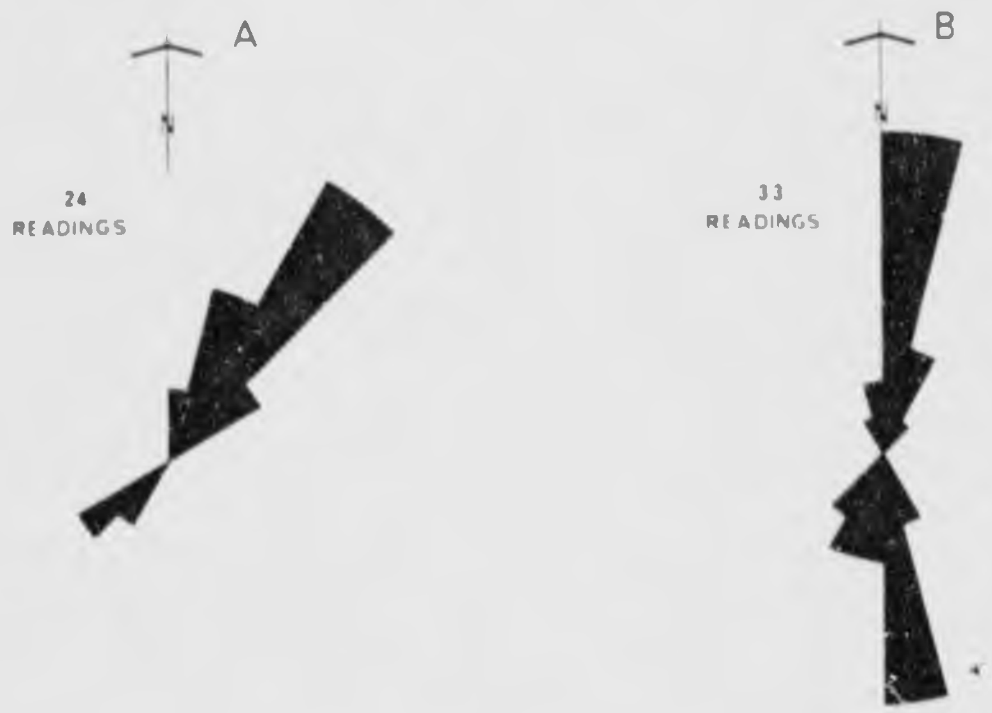

Fig. 31. Cross-bed vector rose diagrams for medium- to coarse-grained sandstone facies: Unit MD3; Saddleback Syncline.

In its lower 80 m this facies consists of laterally persistent planar cross-bedded units with pebble streaks (Fig. 28). Individual cross-bed units average 15 cm, and are up to 30 cm, in thickness. The cross-bedding is generally unimodal (Fig. 29) towards the northeast quadrant, with less common southwesterly oriented foresets (Fig. 31A). On the exposed eastern margin of the sandstone body, however, herringbone cross-bedding (Fig. 30) predominates, resulting in distinct bimodal-bipolar dispersal patterns (Fig. 31B). Other sedimentary structures include reactivation surfaces (Klein, 1970a) and linear ripples. Unusual occurrences of ripple-drift cross-lamination in medium- to coarse-grained sandstones also characterise this facies. Types A and B occur to the exclusion of the in-phase variety. Deformed foreset laminae are rare penecontemporaneous deformation features.

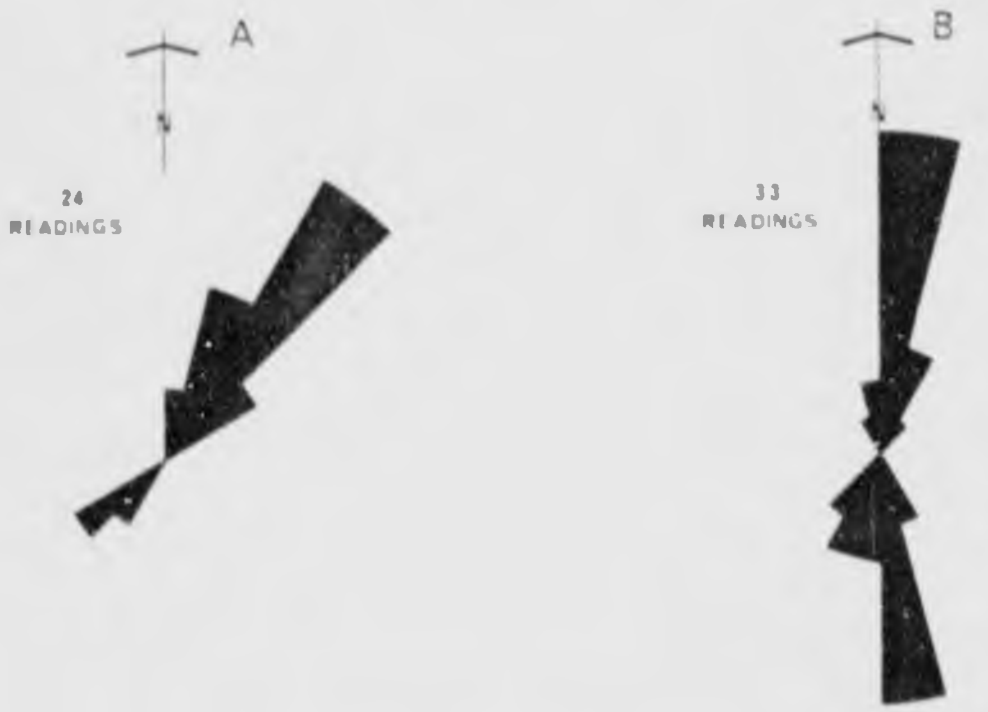

Fig. 31. Cross-bed vector rose diagrams for medium- to coarse-grained sandstone facies: Unit ND3; Saddleback Syncline.

A variant of this facies occurs stratigraphically above the previously described clean sandstone, but separated from it by a 10 m thick unit of interlayered sandstones and shales, analagous to those which enclose the upward-fining cycles of the conglomerate-sandstone facies (Fig. 28). Alternating medium- and coarse-grained subarkoses occur through the upper 40 m thickness (Fig. 28), the coarser variety as thin lenses and more continuous layers up to 15 cm thick. Grains are sub- to well-rounded and sorting is moderate. Chert and polycrystalline quartz are common constituents. Plane-bedding (Fig. 32) is the predominant sedimentary structure, with occasional trough cross-bedding in units less than 8 cm thick. These cross-beds are oriented at right angles to those at the lower level, namely towards the northwesterly quadrant. Ripple-cross-lamination (Fig. 32) is present in the upper part of this unit, where linear ripples are also common. Shale layers, up to 5 cm thick, and thin drapes, are interlayered within the coarser clastics. These are invariably desiccated and often reworked to form discrete angular or curled-up shale-flakes (Figs. 33 and 34). Columnar water-escape structures (Fig. 35; Lowe, 1975) are also present.

B3. Penecontemporaneously Deformed Sandstone Facies.

Fine- to coarse-grained sandstones of this facies are developed through stratigraphic thicknesses upwards of 300 m (Fig. 2) and occur in sharp contact beneath either of the two overlying facies (B1 or B2). The sandstones are felspathic greywackes in composition, are poorly-sorted and

Fig. 32. Plane-bedding and ripple-cross-lamination with shale drape: Unit MD3; Saddleback Syncline (scale in 1 and 5 cm divisions).

Fig. 33. Two-stage desiccation of shale-partings: Unit MD3; Saddleback Syncline (note hammer head for scale).

Fig. 34. [illegible] mud-flakes: Unit MD[illegible], Saddleback Syncline (scale in 1 and [illegible] divisions).

Fig. 35. Columnar weathering structure; Unit MD3; Saddleback Syncline.

grains are moderately rounded to angular. Thin shale-laminae coat 0,5 to 1,5 cm thick (often coarse-tail graded) units which are normally grain-supported at their base and become more loosely-packed within an argillaceous matrix upwards. A single occurrence of desiccation-cracking of a shale-parting was observed.

Primary sedimentary structures are absent from this facies. Soft-sediment or penecontemporaneous deformation structures occur throughout. The best developed of these are water-escape pipes or sheets (Middleton and Hampton, 1973) which may extend vertically for up to 3m (Fig. 36). The shale-coated contacts between graded units are ubiquitously ragged and display small angulate and cupola-shaped irregularities (Fig. 37; Lowe 1975, Fig. 3b) and small-scale flame structures. Silica concretions, up to ½ cm in diameter, frequently occur within the upwarped irregularities between graded units.

iii) Palaeoenvironmental Interpretation.

a) General Environment.

Any depositional model proposed for this facies assemblage must account for the variable lithologies, both in terms of grain-size and textural maturity, and the often bimodal-bipolar pattern of the palaeocurrent vectors. In addition, cognisance must be taken of the upward-coarsening arrangement of the facies assemblage, and the increasing evidence of desiccation upwards. The above criteria imply a depositional environment that was subjected to variable degrees of reworking by tidal processes. An upward-shallowing of the depository with time is also indicated. Prograding

Fig. 36. Water-escape pipes (sheets): Unit MD3; Saddleback Syncline.

Fig. 37. Ragged contacts between graded arenaceous units. Note the small angulate and cupola-shaped irregularities defined by thin clay laminae: Unit MD3; Saddleback Syncline (scale x ½).

marine depositional systems can account for many of these features. Neither the barrier island model (Bernard et al., 1962) nor the Mississippi Delta model (Wright and Coleman, 1974), however, explain all the variables of this facies assemblage. It is intended that each facies be analysed separately before combining them into a single depositional model.

b) Conglomerate-Sandstone Facies (B1)

Upward-fining cycles, analogous to those developed in this facies (Fig. 38) are described from the Rhine and Meuse estuaries of the Netherlands coast (Oomkens and Terwindt, 1960; Terwindt, 1971). Sedimentation in these estuaries is accomplished mainly by the lateral migration of tidal channels which deposit :

1) Coarse-grained, impure, cross-bedded sands overlying a basal lag deposit; and,

2) Medium- to fine-grained, cleaner sands with plane-bedding and abundant thin clay drapes.

The basal lag of the Holocene examples is composed of shell debris, and mud-flakes derived by erosion of channel banks. Shale-flakes are common in the channel lag of the conglomerate-sandstone facies, but the presence of abundant quartz and chert pebbles suggests a strong fluvial input into the depositional system. In this respect, the Moodies cycles are more similar to the tidal channels of the Niger Delta, where Oomkens (1974) has recognized a strong fluvial influence. The tidal channels of the North Sea, by contrast, derive almost all their sediment from offshore (Noordshoorn van der Kruijff and Lagaaij, 1960).

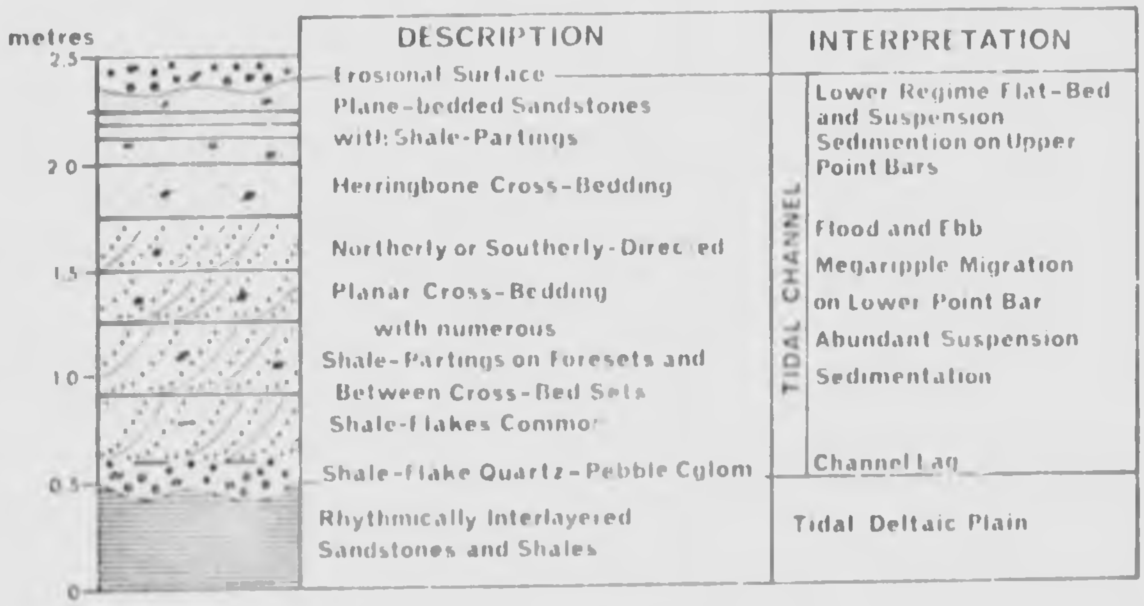

Fig. 38. Inferred depositional environments of conglomerate-sandstone facies.

Coarse-grained, cross-bedded sandstones of this facies rest on the basal lag, and are thought to indicate a transition from upper to lower flow regime conditions from the channel floor to the lower point bar. Ubiquitous planar cross-bedding and small-pebble lenses on the foresets indicate rapid sedimentation in the middle to upper part of the lower flow regime (Simons et al., 1965; Jopling, 1965). Migration of linear or slightly sinous megaripples on lower point bars was both landward and basinward (Fig. 22) in different parts of the estuary. The thin shale laminae commonly present on the foresets of these cross-beds indicate rapidly fluctuating energy conditions. A tidal circulation pattern, with alternating bedload and suspension sedimentation, can be inferred.

The plane-bedded upper unit probably formed on upper point bars in the lower part of the lower flow regime (Klein, 1967b). This is supported by the thin clay laminae distributed throughout the sandstones, which indicate that the critical shear velocity for the initiation of sediment (clay) movement was not surpassed (Terwindt and Breusers, 1973). These were deposited during the turning of the tides, and almost certainly would have been removed by upper flow regime conditions. The associated small-scale cross-bedding also indicates a lower flow regime condition. The bimodal-bipolar character (Fig. 22) of this planar cross-bedding indicates that linear megaripple migration was both landward and basinward. Flood and ebb currents thus appear to have been of equal strength during the deposition of the upper unit of these cycles. Associated ripple-drift cross-lamination probably represents levee or upper point bar deposition. A high suspended to bedload ratio is indicated, with the in-phase variety representing a suspension sedimentation phenomenon under waning energy conditions (Jopling and Walker, 1968). Oomkens (1974) considered that the presence of thin clay laminae on foresets of upward-fining cycles is adequate to distinguish between fluvial and tidal channel-fill sequences. In association with the bimodal cross-bedding these shale laminae thus serve to characterise these cycles as tidal channel deposits.

The channel-fill sediments occur within rhythmically interlayered shales and argillaceous sandstones analagous to the deposits formed under "low-current action" in the Netherlands estuaries (Oomkens and Terwindt, 1960; Terwindt, 1971). They are interpreted as estuarine tidal flat deposits, and in Holocene environments like the Niger Delta (Oomkens, 1974) would probably include marsh deposits. Shale-flakes in the basal lags of this facies were derived from these sediments by lateral migration of tidal channels. These rhythmically layered sediments are also similar to deposits from lower tidal flats of the Wadden Sea, where Van Straaten (1954) described meandering tidal channels cutting across muds and muddy sand deposits.

c) Medium- to Coarse-Grained Sandstone Facies (B2)

In many modern estuaries, particularly where exposed to the open ocean, tidal channel sediments are intimately associated with sand shoals, many of which emerge at low tide. Notable examples are those of the Haringvliet estuary on the Netherlands coast (Oomkens and Terwindt, 1960; Terwindt, 1971) and the Elbe estuary in the German Bight (Reineck and Singh, 1973, p. 318). In both examples extensive reworking of sediments occurs as a result of high tidal ranges.

Many features of the medium- to coarse-grained sandstone facies can be recognized in these modern shoal deposits. The lower 80 m of this facies (Fig. 39) formed largely by linear megaripple migration in the upper part of the lower flow regime. Such bedforms predominate on the surface of the Elbe shoals, and are most abundant below low-water level (Dörjes, et al., 1970). The dominant ebb orientation (Fig. 31A) of planar cross-bedding through much of this facies indicates that offshore currents washed over the surface of the shoals. Strong offshore-directed currents have been recorded from the German Bight (Gadow and Reineck, 1969)

Such currents may be capable of transporting pebbles, and could account for the conglomerate lenses in this facies, which appear to occupy tidal scours. The pebbles were probably derived from nearby estuaries. Furthermore the ripple-drift cross-lamination often developed in these sandstones, points to currents capable of transporting a significant quantity of medium- and coarse-grained sands in suspension. On the eastern extremity of the sandstone body abundant bimodal-bipolar cross-bedding (Fig. 31B) suggests that both flood and ebb currents operated on the margins of the shoal, clearly illustrating the time-velocity asymmetry of the tidal currents (Klein, 1970a). Rare reactivation surfaces lend further support to the contention that this facies accumulated in a tide-dominated environment.

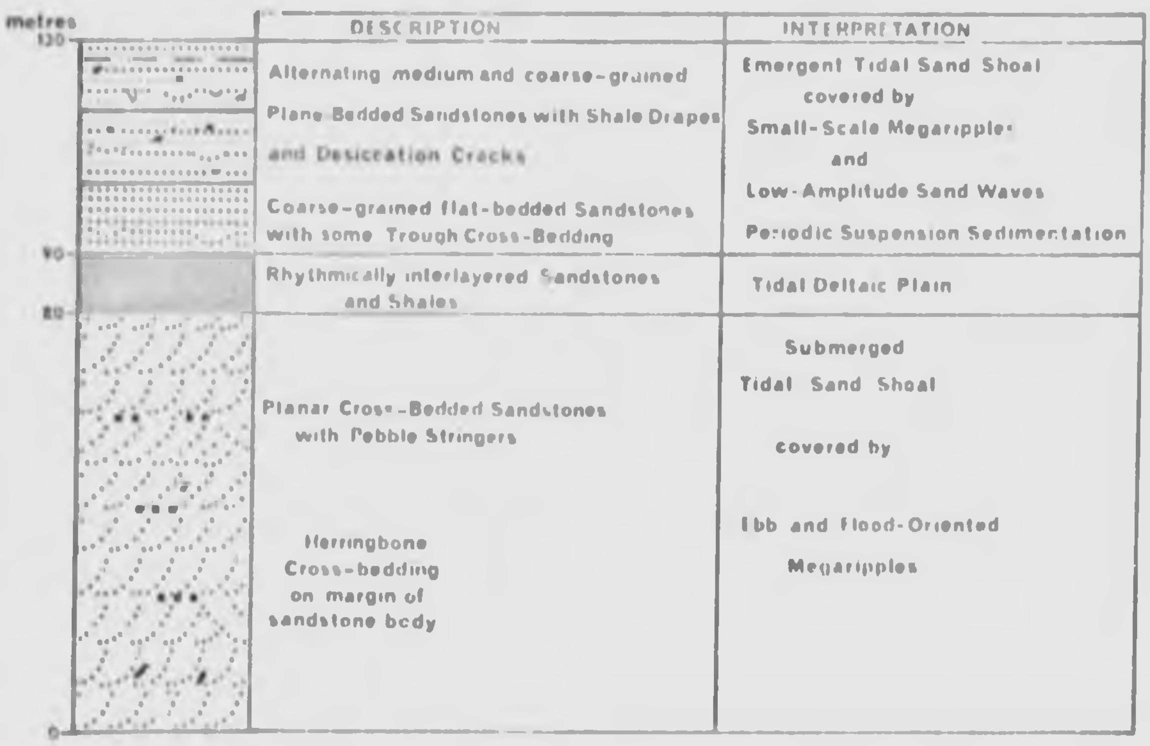

Fig. 39. Inferred depositional environments of medium- to coarse-grained sandstone facies.

On the Elbe shoals, megaripples are less common above wave-base, and plane-bedded, coarse-grained sands, displaying current ripples, predominate. At still shallower depths, and especially above low-water line, muds are associated with sands (Dörjes, et al., op cit.). The upper 10 m of the medium- to coarse-grained sandstone facies (Fig. 39) display characteristics similar to the shallow-water portions of the Elbe shoals. Plane-bedded sandstones and small trough cross-beds, with shale flasers and desiccation cracks in the upper parts, indicate a gradual shallowing. The term "tidal bedding" has been applied by Wunderlich (1970) to alternating sands and muds of this type. Structures within the sands, such as ripple-cross-lamination (Fig. 32), suggest deposition in the lower part of the lower flow regime. Interlayered, horizontally-laminated, coarse- and medium-grained sands are diagnostic of this regime (Jopling, 1964) and may form by the migration of very low amplitude sand waves (Smith, 1971).

In summary, it is proposed that the lower and upper sand-dominated units of this facies accumulated below and above wave base respectively. The rhythmically laminated shales and sandstones separating these two units (Fig. 39) are identical to the tidal deltaic plain deposits in which the tidal channel sequences occur, and represent a lens of this facies sandwiched within a prograding shoal assemblage.

d) <u>Penecontemporaneously Deformed Sandstone Facies (B3)</u>

The lack of preserved primary sedimentary structures and the immature, often loosely packed nature of the sediments in this facies indicate rapid deposition in the absence of bedload processes. The frequently encountered coarse-tail grading of depositional units is suggestive of gravity flow processes (Middleton and Hampton, 1973). One possible gravity flow mechanism is liquefaction, which occurs when an unconsolidated grain-supported sediment is transformed into a transitory fluid-supported suspension as a result of a rapid but temporary increase in the pore fluid pressure (Lowe, 1975). Sedimentary particles are transported in this state, in response to gravity, as long as the pore pressure exceeds the hydrostatic pressure. Resedimentation (upon loss of pore pressure) results in the downward settling of solids through the fluid (Lowe, 1976). This often develops basal grain-supported coarse fractions similar to the coarse tails present within individual depositional units of this facies. Deposition of the finer-grained debris from a suspended state in diluted turbidity currents (Lowe, 1976) could account for the upper loosely-packed portions of graded units, and the capping shale laminae.

Penecontemporaneous deformation structures in liquefied flow deposits form as a result of water migration either during or after deposition of overlying units (Lowe, 1976). The vertical persistence of water-escape pipes or sheets in this facies points to the simultaneous dewatering of depositional sequences upwards of 3 m in thickness. The ragged contacts between individual depositional units (Fig. 37), and specifically the small-scale flame structures, probably formed through loading and/or passive, as opposed to forceful, water escape. Evidence for the accumulation of saturated waters below the thin shale laminae is provided by the common silica concretions within the upwarped irregularities. The lack of dish and pillar structures (Lowe and Lo Piccolo, 1974) in this facies suggests that rapid sedimentation resulted in an excessive load, which limited the forceful escape of water to large cross-cutting pipes and sheets which may follow penecontemporaneous joints.

Liquefied sediment flow deposits are not diagnostic of specific sedimentary environments. Although rapidly deposited sediments and their associated dewatering structures most commonly form in deep-waters, they may also develop in delta front and alluvial fan depositional environments (Lowe and Lo Piccolo, 1974). The association of this facies with delta plain and sand shoal deposits suggests a delta front as the most likely depositional setting. Repetitive liquefied gravity flow surges on a steepened slope fronting the delta plain could adequately account for the primary and secondary characteristics of this facies. Rare subaerial exposure of the upper part of the delta front, where in contact with the delta plain to the exclusion of the sand shoal facies, would have resulted in occasional desiccation of shale laminae within this facies.

e) Discussion.

At first appearance the upward-coarsening sequence of this facies assemblage displays few features in common with modern-day deltas. Often, however, deltas which experience high tidal ranges display primarily tidal flat depositional characteristics, and differ in most respects from deltas that develop under lower tidal ranges, or which are subjected to high wave energies (see Fig. 40). Tidal processes at river mouths produce intensive meandering above the zone of tidal influence, cause tidal channels to be sand-filled, and result in the accumulation of large, linear sandy tidal ridges seaward of the river mouth (Coleman and Wright, 1975). The Ord River Delta in northern Australia, illustrates the above processes. Semi-diurnal tides have a mean range of 3,80 m, and an average spring range of 5,15 m, but an upstream amplification of the tidal wave causes mean and spring tidal ranges of 4,75 and 6,60 m, respectively, within the delta itself. Tidal currents are orientated dominantly in an offshore-onshore direction, and, as is also the case for the macrotidal Fly Delta (Galloway, 1975; Fig. 40) produce sand shoals, covered by megaripples, oriented perpendicular to the shoreline (Wright et al., 1975).

The three facies of assemblage 'B' can be readily reconciled with a deltaic depositional model of high tidal energies. The penecontemporaneously deformed sandstones (B3) of delta front origin were prograded across by elongate sand shoal (B2) and deltaic plain, including tidal channel (B1), accumulations.

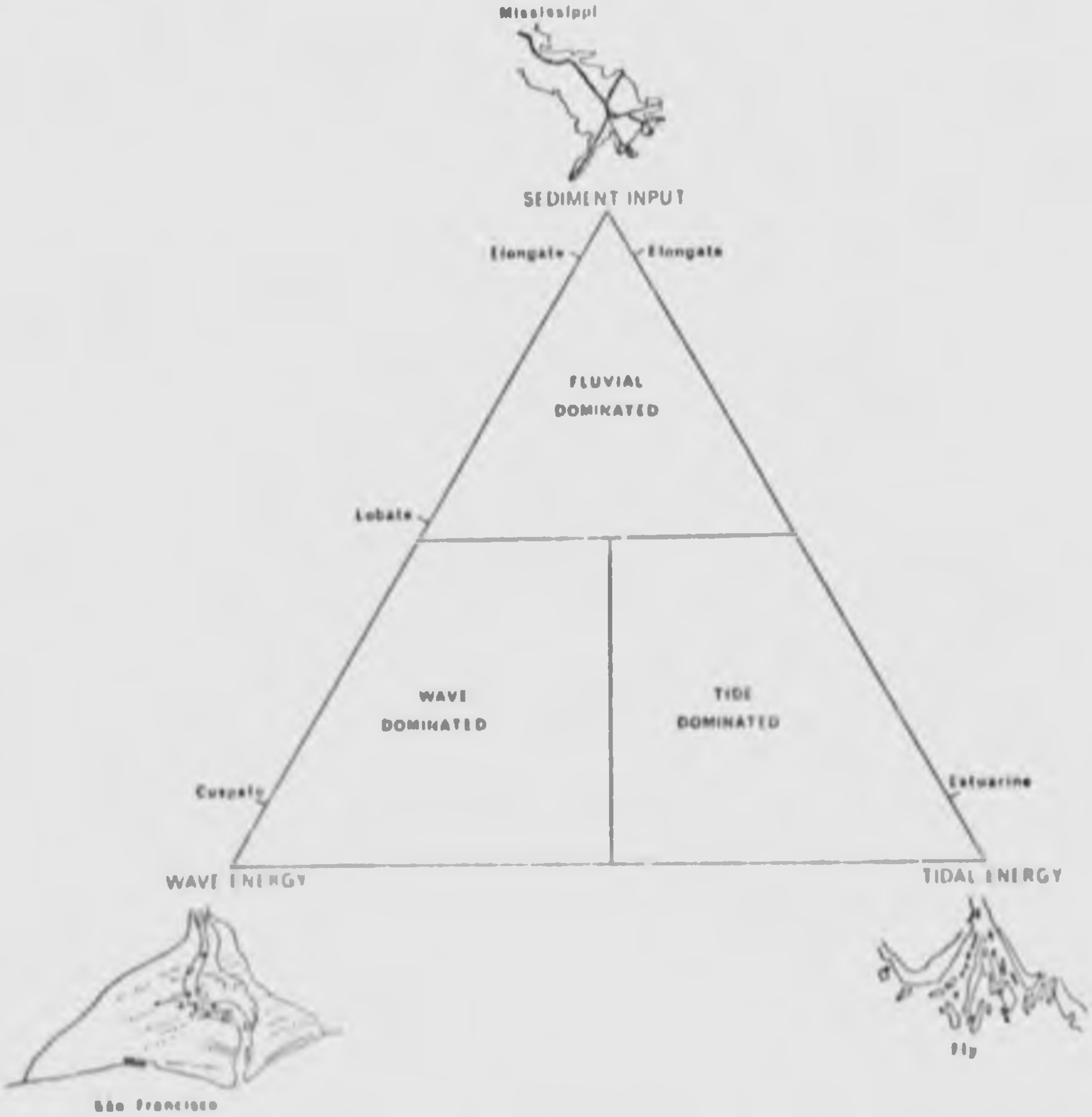

Fig. 40. Classification of deltas according to dominant depositional process (modified after Galloway, 1975).

DEPOSITIONAL FACIES ASSEMBLAGE 'C'

i) Introduction.

Where fully-developed, this assemblage contains four lithologic facies arranged in upward-coarsening stratigraphic sequences. The sequence commences with a thick pile of fine-grained sediment, capped by cross-bedded and plane-bedded orthoquartzitic sandstones (Fig. 41) and is best developed in unit MD3 in the Eureka and Stolzburg synclines (Fig. 2). The overlying stratigraphic unit in the same synclines (MD4) contains the two finer-grained facies of this assemblage, but the orthoquartzitic sandstones are preserved only in the Eureka Syncline, where represented by the plane-bedded facies at the top of unit MD4. The upper orthoquartzitic sandstones of units MD1 and MD3 in the Saddleback Syncline (Fig. 2) are also of the plane-bedded variety. Conglomerates are associated with the upper orthoquartzitic sandstone of unit MD3 in the Saddleback Syncline, and are intercalated on a fine scale with similar sandstones at the top of unit MD4 in the Stolzburg Syncline (Fig. 2).

ii) Description of Facies.

C1 Siltstone - Shale - Banded Iron Formation Facies.

Banded iron formations, up to 20 m in thickness, and consisting of alternating jasper and hematite layers, occur at the base of this facies assemblage (Fig. 41). These are mainly horizontally-laminated with less common poddy-bedding (Fig. 42). The overlying shales (Fig. 41), which are also horizontally-laminated, contain intercalated jasper lenses, laterally-persistent jasper layers averaging 4 cm in thickness, and siltstone lenses. The proportion of siltstone alternating with shale increases upwards, until the upper 50 m contains roughly equal proportions of the two lithologies (Fig. 41). Sedimentary structures in the siltstones are horizontal-lamination and ripple-cross-lamination.

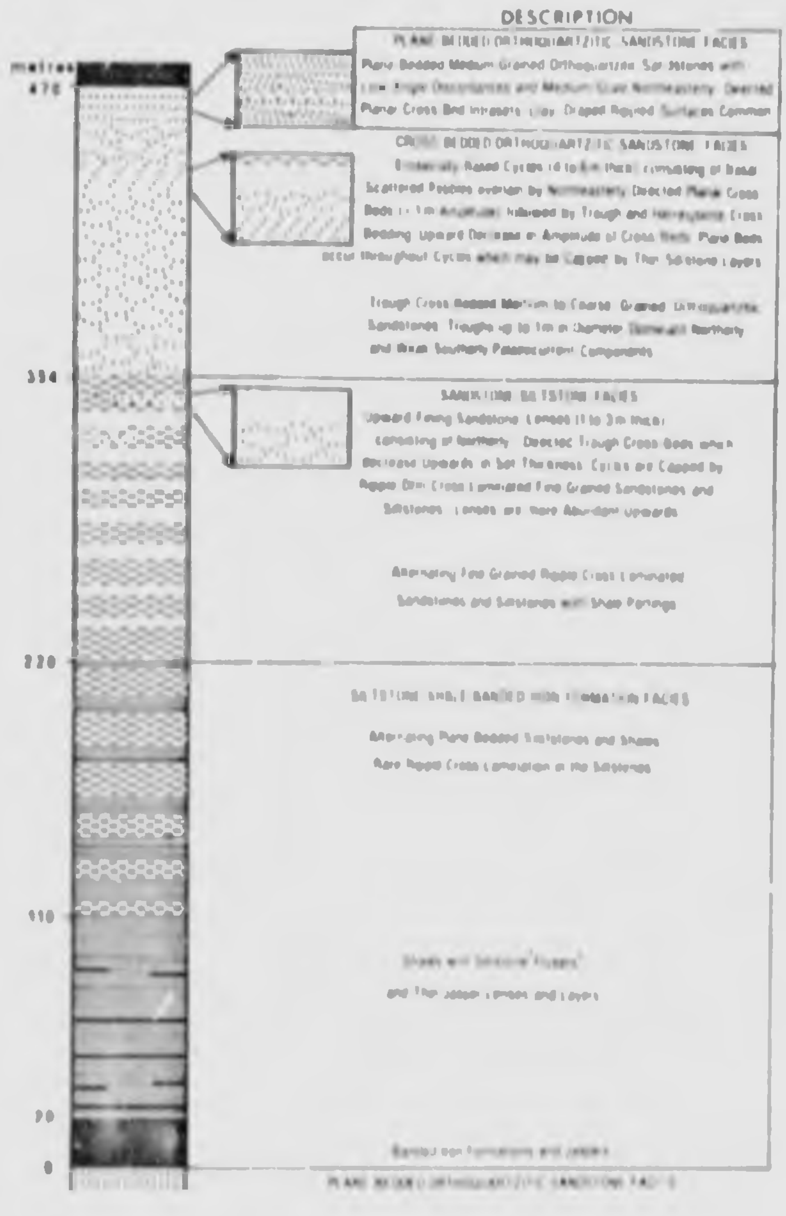

Fig. 41. Measured stratigraphic section through depositional facies assemblage 'C' : Unit MD3; Eureka Syncline.

C2 Sandstone - Siltstone Facies.

This facies has a gradational contact with facies C1 and is composed almost exclusively of interlayered fine-

Fig. 42. Banded iron formation consisting of interlayered hematite and jasper: Unit MD3; Stolzburg Syncline (scale in 1 and 5 cm divisions).

Fig. 43. Alternating fine-grained sandstones and siltstones with shale partings: Unit MD3; Eureka Syncline.

Fig. 44. Overlapping trough cross-beds; sandstone-siltstone facies: Unit MD3; Eureka Syncline.

Fig. 45. Herringbone cross-bedding; cross-bedded orthoquartzitic sandstone facies: Unit MD3; Eureka Syncline.

grained sandstones and siltstones with thin shale layers (Fig. 43). The sandstones and siltstones consist of well-rounded quartz grains in an argillaceous matrix and be classified as quartzwackes. Sedimentary structures in he wackes are dominantly ripple-cross-lamination. Upward-fining quartzwacke lenses are commonly developed in the upper 50 m of the facies. These vary from 1 to 3 m in thickness, 10 to 20 m in strike length and increase in size and abundance upwards. Medium-grained sandstones at the base are structured by northerly-directed small-scale trough cross-beds (Figs. 44 and 46B) and pass upwards into fine-grained sandstones and siltstones containing types A and B, and in-phase ripple-drift cross-lamination (Fig. 41).

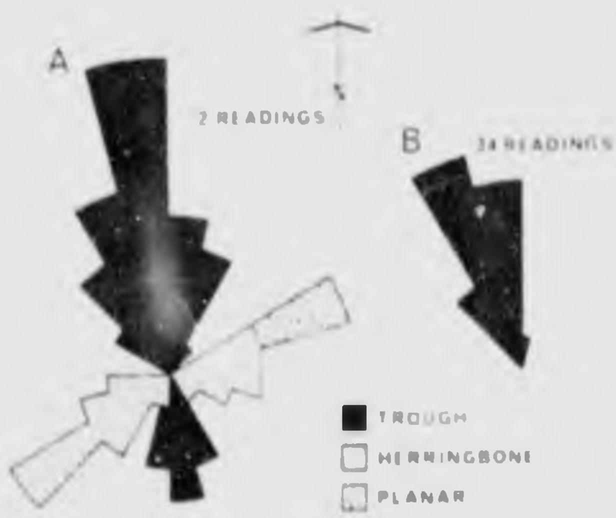

Fig. 46. Rose diagrams for cross-bedded orthoquartzitic sandstone facies (A) and upward-fining lenses within sandstone-siltstone facies (B): Unit MD3; Eureka Syncline.

C3 Cross-Bedded Orthoquartzitic Sandstone Facies.

Cross-bedded orthoquartzitic sandstones are laterally persistent around the Eureka and Stolzburg synclines (MD3) with little variation in stratigraphic thickness, and rest with a sharp contact on the underlying sandstones and siltstones (Fig. 41). Well-rounded, medium- to very coarse-grained and often granular quartz arenites are the predominant lithologies in the lower two-thirds of the facies. Overlapping small- and medium-scale trough cross-beds constitute the only sedimentary structures. These are directed mainly to the north but with a weak southerly mode also present (Fig. 46A). Cross-beds decrease upwards in thickness from 20 to 10 cm through units of up to 3 m.

The upper 25 m of this facies consists of stacked 4 - 6 m upward-fining cycles (Fig. 41). Where fully-developed these contain thin basal conglomerates, consisting of small scattered quartz and chert pebbles, passing upwards into fine- and medium-grained quartz arenites structured by medium- and large-scale planar cross-beds directed to the northeast (Fig. 46A). Herringbone cross-beds (Fig. 45), with a northeast and southwest bimodal-bipolar orientation (Fig. 46A), are frequently developed towards the top of the cycles which may be capped by 10 to 30 cm thick horizontally-laminated and ripple-cross-laminated siltstone layers. Plane-bedded sandstones are scattered throughout the cycles and shale drapes are common along foreset surfaces. The upward-fining cycles are seldom preserved in their entirety with the herringbone cross-bedded and silty units frequently absent.

C4 Plane-bedded Orthoquartzitic Sandstone Facies.

The orthoquartzitic sandstones of this facies (Fig.41) contain between 75 and 90 percent quartz. Well- to moderately-rounded, medium-grained and generally well-sorted quartz arenites predominate but sublitharenites and subarkoses, which may contain up to 10 percent matrix, are also common. Where capping upward-coarsening sequences, the orthoquartzitic sandstones are laterally persistent but vary in thickness along strike from 10 to 20 metres. Flat, lensoid bodies are more common where this facies is enclosed within impure sandstones (see Fig. 2). Cumulative curves (Fig. 47) and statistical parameters (Table 4) for samples from one of these sandstones indicate their high degree of maturity.

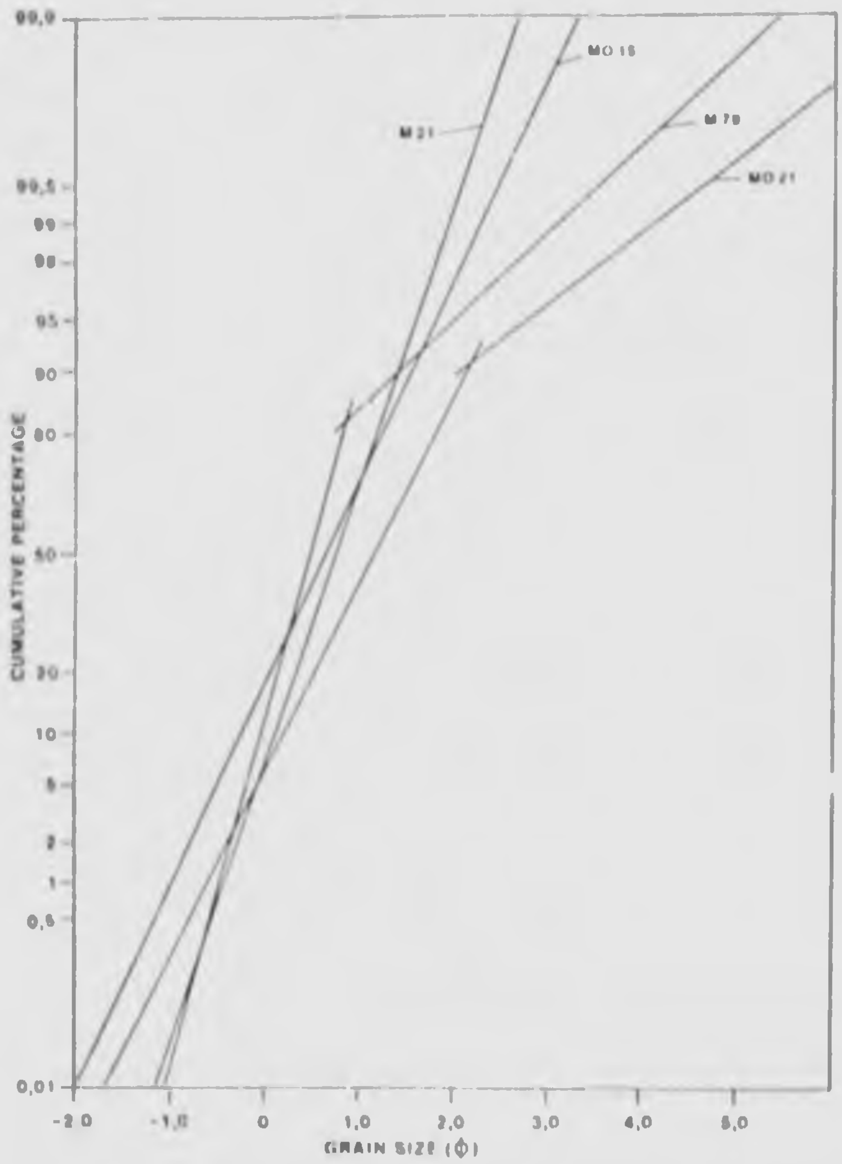

Fig. 47. Grain-size cumulative frequency curves for plane-bedded orthoquartzites at top of unit MD5, Saddleback Syncline. Curves drawn for converted sieve equivalent percentiles.

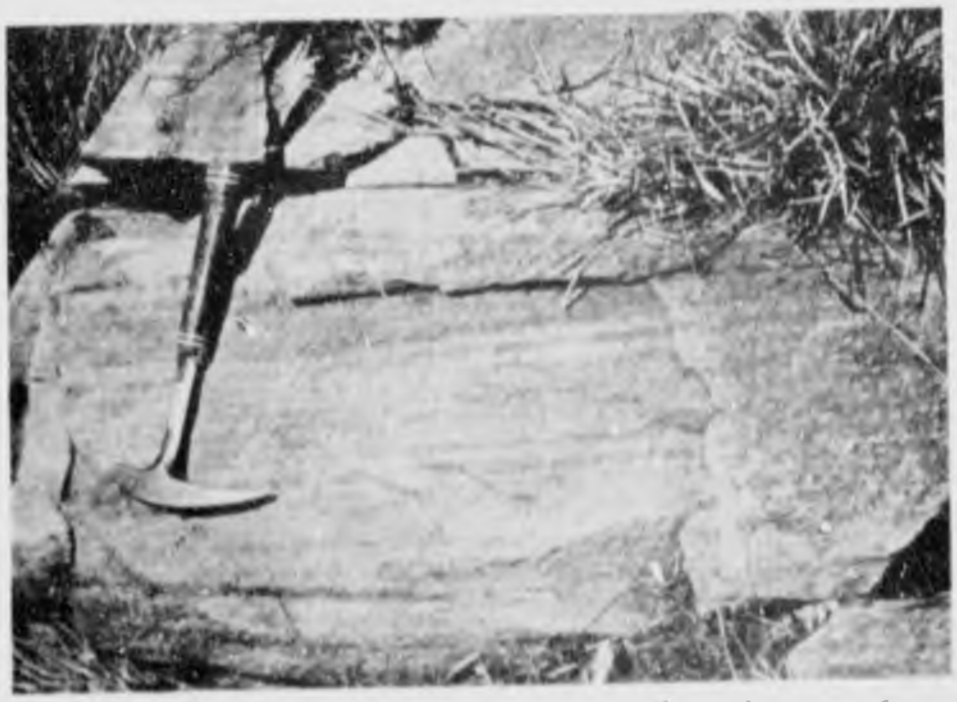

Fig. 48. Plane-bedding with low-angle discordances; plane-bedded orthoquartzitic sandstone facies: Unit MD3; Eureka Syncline.

Fig. 49. Planar cross-bed intraset; plane-bedded orthoquartzitic sandstone facies: Unit MD3; Eureka Syncline.

TABLE 4

STATISTICAL PARAMETERS FOR MOODIES ORTHOQUARTZITES

INTERCALATED WITHIN 'IMPURE SANDSTONES'

Sample No.	M21	MO15	MO21	M79
Mean (Ø)	0.77	0.67	1.18	0.54
Variance (Ø)	0.28	0.52	0.56	0.21
Std. Dev. (Ø)	0.53	0.72	0.75	0.46
Number of Measurements	200	200	200	200

All samples from top of Unit MD3, Saddleback Syncline. (see Figure 2 for stratigraphic position).

Procedures for calculating the statistical parameters were the same as discussed in Table 3.

The orthoquartzitic sandstones are mainly plane-bedded (Fig. 48) with low-angle discordances. Occasional heavy mineral layers occur. Planar cross-bed intrasets (Fig. 49) vary in thickness from 15 to 80 cm and are easterly-directed in all instances. Where capping upward-coarsening stratigraphic sequences, the orthoquartzites often contain linguoid, sinuous, linear and weakly-developed interference rippled surfaces and occasional washout ripples. These are invariably coated by mm-thick shale-partings, which are occasionally desiccated. Vertical sequences of sedimentary structures, through 3 to 5 m thick cycles, consist of planar cross-beds overlain by plane-bedding with rippled surfaces becoming more abundant upwards. Poorly-packed conglomerates, in vertically discrete beds, occur within or on top of some of the lensoid orthoquartzitic sandstone bodies.

iii) Palaeoenvironmental Interpretation.

a) General Environment.

The maturity of the arenaceous sediments of this facies assemblage coupled with the complex palaeocurrent patterns (Fig. 45) are suggestive of marine processes. The upward-coarsening arrangement of the facies (Fig. 50) supports this contention, and is compatible with the prograding barrier island-shallow shelf depositional models of Bernard et al. (1962), Howard and Reineck (1972) and Reineck and Singh (1971). Certain features are, however, at variance with these models, such as the upward-fining cycles in facies 'C3' and the abundant shale drapes associated with facies 'C4'. For these reasons the depositional processes responsible for each facies will be considered separately before developing a single depositional model for this specific barrier island sequence.

b) Siltstone-Shale-Banded Iron Formation Facies (C1)

The abundant banded iron formations and shales in the lower half of this facies (Fig. 50) indicate quiet water chemical and suspension sedimentation on an offshore shelf. Banded iron formations were precipitated in environments furthest removed from terrigenous input, and are the deepest-water or most distal sediments of this facies assemblage. Finely-alternating ferruginous and siliceous layers in banded iron formations most readily form as a response to cyclicly fluctuating Eh and pH conditions (Walter, 1972). Photosynthetic organisms are the agents most likely to have effected such short-term changes, be they diurnal or seasonal, and it is possible that primitive blue-green algae inhabited shallow photic shelves during the Archaean. Although reflecting different depositional processes, the laterally persistent shelf shell beds of the Mississippi Delta (Coleman and Gagliano, 1965) and the Pennsylvanian marine platform carbonates of the Eastern Interior Basin (Pryor and Sable, 1974) both of which represent intervals of non-detrital deposition, are possible analogues of the chemical sediments of this facies. The overlying shales with jasper lenses and layers indicate an offshore environment, intermittently receiving fine-grained detritus which settled from suspension and alternated with short periods of siliceous chemical sedimentation at relatively low pH's and high Eh's. At shallower water depths, a continuous supply of fine-grained detritus prevented the formation of chemical precipitates. Dominant suspension sedimentation and periodic migration of 'starved' ripples developed shales with siltstone lenses. Low-energy bedload processes became increasingly significant with progressive shallowing of the offshore shelf resulting in interlayered, laterally-persistent ripple-cross-laminated and horizontally-laminated siltstones and shales.

Fig. 60. Palaeoenvironmental interpretation for depositional facies assemblage 'C'.

c) Sandstone-Siltstone Facies (C2)

Lithologies and sedimentary structures developed through much of this facies are analogous with those reported from the nearshore shelf off Sapelo Island (Howard and Reineck, 1972), and indicate that low-energy bedload and infrequent suspension sedimentation were the dominant depositional processes. The common coarser-grained, upward-fining lenses developed towards the top of the facies (Fig. 50), however, imply periodic higher- but waning-energy currents on the nearshore shelf. These currents had a predominant northerly or offshore orientation, as indicated by the cross-bed rose diagram for these lenses (Fig. 46B), and were probably generated by rip currents, which operate perpendicular to modern coasts. As a result of longshore processes and onshore mass transport by incident waves, water accumulates inside breaker zones (Huntley and Bowen, 1975; Komar, 1975). The seaward escape of this water sets up offshore currents capable of erosion and transportation of sand through the upper shoreface (Davis and Fox, 1972; Vos, 1976). Broadening of the rip current into a rip head results in waning-flow (Huntley and Bowen, op cit.) which generates conditions suitable for the offshore migration of megaripples, and the formation of trough cross-beds. A further decrease in velocity results in current ripple migration and suspension sedimentation of fine-grained particles. Ripple-cross-laminated and ripple-drift cross-laminated fine-grained sandstones and siltstones are developed, and represent decreasing bedload to suspended load deposition. Upward-fining cycles similar to those present in this facies, and recording single depositional events, are generated in this way.

d) Cross-bedded Orthoquartzitic Sandstone Facies (C3)

In progradational barrier island sequences, shelf deposits most commonly coarsen gradationally upwards into shoreface sands. The sharp basal contact (Fig. 50) and the dominant offshore palaeocurrent component for trough cross-beds of this facies (Fig. 46A) are not typical of normal shoreface sediments. A possible analogue exists in the Miocene of California, where Clifton (1973a) has attributed shoreface sediments with a sharp basal contact to channelized rip currents which reworked sediment within the zone of wave build-up and surf. Resurgent rip currents operating in the lower shoreface are invoked as being responsible for the lower two-thirds of this facies (Fig. 50), with the upward-decrease in trough cross-bed thicknesses as the only manifestation of waning rip current velocities. The weak onshore palaeocurrent component (Fig. 46A) can be related to the shoreward motion of shoaling waves (Clifton, et al., 1971; Greenwood and Davidson-Arnot, 1975) between individual rip current pulses.

The upward-fining cycles developed in the top 25 m of this facies (Fig. 50) closely resemble tidal inlet deposits from Long Island (Kumar and Sanders, 1974). Tidal inlets are narrow channels between barrier islands dominated by tidal currents whose formation has been attributed by Pierce (1970) to tidal surges associated with storms, which cause water to flow over, and possibly incise into, barrier islands. Depending on the magnitude of longshore currents, tidal inlets vary in orientation ranging from perpendicular to acute angles to the shoreline (Kumar and Sanders, op cit; Hubbard and Barwis, 1976). The modal migration direction of ebb-oriented bedforms in South Carolina inlets is more commonly subparallel rather than perpendicular to the shoreline (Hubbard and Barwis, op cit.). Tidal inlets migrate parallel to longshore currents, with resultant spit development on the up-current end of the inlet. Hydrodynamic conditions vary in different parts of inlets and develop characteristic sequences of sedimentary structures. Basal conglomerates in the cycles of this facies are interpreted as channel floor deposits. Overlying planar cross-beds were generated by migration of ebb-oriented sand waves in deep channel portions of tidal inlets. The interlayered plane-bedded sandstones are thought to have developed during upper flow regime conditions, probably as a response to periodic water-level changes. Whereas deeper parts of tidal inlets are generally characterised by unidirectional flow, shallow-water realms frequently contain bidirectional currents related to ebb and flood tides (Van Beeck and Koster, 1972). The herringbone cross-bedding at the top of the cycles can be related to such processes, while the common shale-drapes on cross-bed foresets provide further evidence of a strong tidal influence during the deposition of these sediments (Oomkens, 1974). Palaeocurrent patterns for these cycles (Fig. 46A) are compatible with tidal inlets oriented in a northeast-southwest direction, which is subparallel to the palaeoshoreline. This, by implication, indicates a strong easterly longshore drift component. The vertical superimposition of tidal inlet cycles is due to successive longshore migration and reopening of inlets at their original sites (Kumar and Sanders, op. cit.). Extensive scouring during longshore migration results in incomplete preservation of underlying inlet sequences.

e) Plane-bedded Orthoquartzitic Sandstone Facies (C4)

The sediments of this facies (Fig. 50) superficially resemble foreshore deposits with the plane-bedded sandstones developed by swash processes and the planar cross-bed intrasets representing ridge and runnel accumulations (Hayes and Boothroyd, 1969; Wunderlich, 1972). Abundant clay-draped ripple marks are, however, atypical of foreshores, while the planar cross-beds have a longshore attitude, which contrasts with the normal landward-dip of ridge and runnel cross-stratification. A depositional environment influenced by swash, longshore currents and incorporating protected settings suitable for suspension sedimentation, is indicated. Barrier spits contain subenvironments in which each of the lithologies of this facies could have accumulated. Subaqueous spit-platforms consist of long, planar cross-beds generated by beach drifting in response to longshore currents (Meistrel, 1972; Kumar and Sanders, 1974). Spit beach faces are subjected to upper flow regime swash processes, and are structured by plane-bedded sandstones with low-angle discordances. Bays commonly occur behind the spit beachface, and are conducive to current ripple migration, modification of rippled surfaces during falling water-level, and periodic suspension sedimentation. The sequences of lithologies and sedimentary structures developed in this facies can be accounted for through seaward progradation of these three subenvironments.

Where enclosed within impure sandstones (See Fig. 2), the dominance of plane-bedding, the well-sorted nature of the sandstones as reflected in the shape of the cumulative frequency curves, (Fig. 47. Visher, 1969), and the less common planar cross-bed intrasets in this facies indicate swash and longshore drift as the dominant depositional processes. The associated conglomerates at the top of unit MD3 in the Saddleback Syncline and the top of unit MD4 in the Stolzburg Syncline (Fig. 2) are similar to wave-worked conglomerates described by Clifton (1973b). Their vertically discrete nature and intercalation with clean sandstones serves to distinguish them from alluvial gravels.

f) Discussion.

The morphology of Holocene coastlines is determined to a large extent by tidal ranges (Hayes, 1975). Mesotidal coasts are characterised by short and stubby barrier islands and are dominated by tidal deltas. Barrier islands along microtidal coasts, by contrast, are markedly elongated and, because of the dominant wave effects, contain features such as spits and washover fans. Tidal currents are only important at inlet mouths. The abundant spit deposits inferred in the upper facies of this assemblage, coupled with the absence of ebb tidal delta accumulations, is suggestive of a microtidal coastline similar to that fronting Laguna Madre in Texas (Dickinson, et al., 1972). Furthermore, rip currents only occur where high wave conditions exist, while deposits formed by these currents are most commonly preserved along microtidal coastlines free from the modifying effects of tidal processes (Vos and Hobday, in press). Tidal currents appear to have been confined to tidal inlets during the development of this facies assemblage.

DEPOSITIONAL FACIES ASSEMBLAGE 'D'

i) Introduction.

Two facies are distinguishable in this assemblage and alternate through stratigraphic unit MD2 in the Eureka and Stolzburg synclines (Fig. 2). These facies also occur in unit MD5, as 5 to 10 m thick sequences interlayered within the impure sandstone-siltstone-shale facies (A3; Fig. 2). Argillaceous sandstones and siltstones with abundant shale-partings, and shales with sandstone and siltstone layers and lenses characterise this facies assemblage.

ii) Description of Facies.

D1. Sandstone - Shale Facies.

This facies is characterised by upward-fining cycles, from 1,5 to 6 m in thickness (Fig. 51A). A basal sandstone is generally less than 1 m thick, and is a fine-grained subarkose, often massive but more commonly avalanche herringbone cross-bedded with bimodal-bipolar palaeocurrent patterns (Fig. 52A). Complex internal arrangement of bedding (Fig. 53) and occasional reactivation surfaces, are also common in these basal sandstones. Cross-bedded units in the upper parts of the sandstones often contain shale flasers. Flaser bedding, with associated rill marks, becomes more distinct upwards, where quartzwackes contain thin shale lenses and drapes (Fig. 54). Ripples are common, and include interference (Fig. 55), linear, lunate and linguoid varieties. A rose diagram plot shows that the ripple crests are apparently randomly oriented with respect to the strike of the underlying herringbone foresets (Fig.52A).

Wavy bedding is commonly developed upwards in the cycles, which are terminated by ubiquitous lenticular bedding. Different types of lenticular bedding occur, including connected types with thick (Fig. 56) and with thin (Fig. 57) sandstone lenses (Reineck and Wunderlich, 1968). Shales in the lenticular units often display desiccation, with some reworking of the clasts to form shale-flake conglomerates.

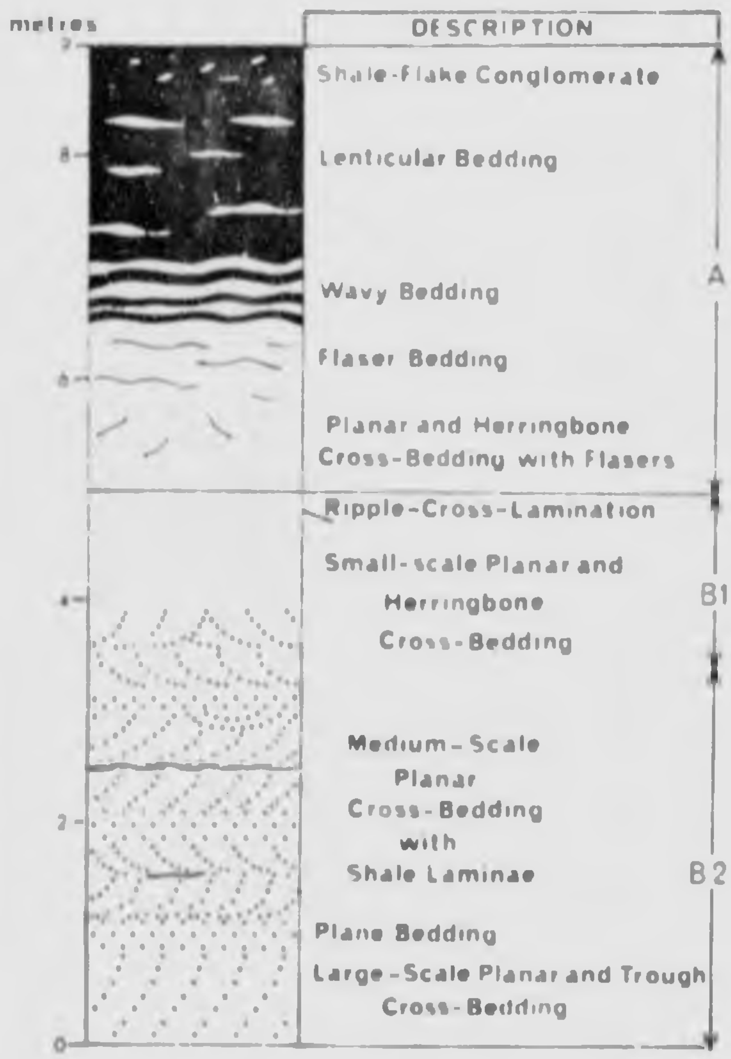

Fig. 61. Idealised stratigraphic column through sandstone-shale (A) and fine- to medium-grained sandstone (B) facies; Unit MD1; Eureka Syncline.

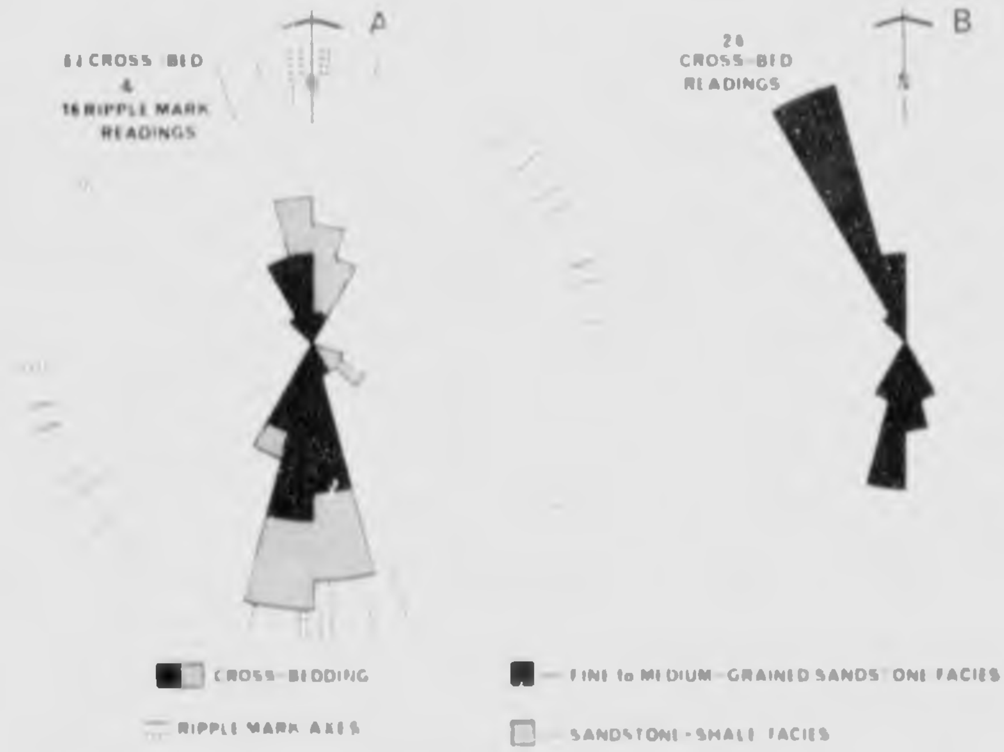

Fig. 52. Cross-bed vector rose diagrams for sandstone-shale and fine- to medium-grained sandstone facies. A: Small-scale cross-beds and ripple marks; B: Medium- and large-scale cross beds: Unit MD1; Eureka Syncline.

Fig. 53. Complex internal organization of bedding: Unit MD2; Eureka Syncline (scale in 1 and 5 cm divisions).

Fig. 54. Flaser bedding: Unit MD2; Eureka Syncline(scale in 1 cm divisions).

Fig. 55. Interference ripple marks: Unit MD2; Eureka Syncline (note hammer for scale in top right hand corner).

Fig. 56. Lenticular bedding with thick sandstone lenses: Unit MD2; Eureka Syncline (scale in 1 and 5 cm divisions).

Fig. 57. Lenticular bedding with thin sandstone lenses: Unit MD2; Eureka Syncline (scale in 1 cm divisions).

D2. Fine- to Medium-Grained Sandstone Facies.

Fine- to medium-grained subarkoses and quartz wackes alternate with the sandstone-shale facies through successions of almost 1000 and 600 m in the Eureka and Stolzburg synclines, respectively (Fig. 2). Both sandstone varieties contain appreciable amounts of potash and plagioclase felspars, are devoid of chert, and contain only rare polycrystalline quartz grains. Where not obscured by authigenic overgrowth, grains are rounded to subangular, with the subarkoses tending to be better sorted than the quartz wackes. In the Eureka Syncline, biotite is the predominant matrix constituent of the wackes, having formed by metamorphism of clays. Clay minerals are still preserved in wackes from the Stolzburg Syncline.

Different associations of sedimentary structures enable two subfacies to be recognized in the fine- to medium-grained sandstones. The first, which tends to be more argillaceous than the second, consists of small-scale planar and herringbone cross-bedding (Fig. 52A) capped by ripple-cross-lamination. Linear ripples are oriented parallel to the strike of the cross-bed foresets. The second subfacies is intimately associated with, and often gradational into the first. Medium- and large-scale planar cross-bedding (Fig. 58), as well as plane-bedding, characterise this subfacies (Fig. 51, B2). Bimodal- bipolar palaeocurrent patterns are again present (Fig. 52B). Penecontemporaneous deformation structures include overturned foresets and convolute lamination (Fig. 59). A variant of the second subfacies in the Stolzburg Syncline

Fig. 58. Large-scale planar cross-bedding : Unit MD2; Eureka Syncline.

Fig. 59. Convolute lamination in medium-grained sandstones: Unit MD2; Eureka Syncline.

consists of lenticular, cross-bedded sandstones separated by 2 to 5 cm thick shale- and siltstone-partings and overlain by 5 to 8 m thick sequences of plane-bedded and low angle cross-bedded sandstones.

iii) Palaeoenvironmental interpretation.

a) General Environment.

Evidence from the above two facies suggests a depositional environment influenced by tidal processes. The ubiquitous bidirectional palaeocurrent patterns are indicative of bedload transport with bipolar reversals of flow direction. This phenomenon is most commonly associated with flood and ebb tidal currents. Interbedded sandstones and shales point to alternating bedload and suspension sedimentation. Mud deposition in a marine environment occurs in subtidal regimes below wave base, or on tidal flats where low wave and current energies prevail (Reineck and Singh, 1973, p. 358). The common occurrence of shale-flakes and desiccated shale drapes supports a tidal flat environment of alternating inundation and exposure. The low energy condition required for suspension sedimentation occurs at slack high water.

b) Sandstone - Shale Facies (D1)

On extensive tidal flats, such as along the north German and Netherlands coasts, fine-grained sediments accumulate near the high-water line and coarser sandy sediments around low-water level (Reineck, 1967; Van Straaten and Kuenen, 1957). On the mid-flat, interbedded sands and muds are developed. The sandstone-shale facies (Fig. 60A) can best be equated with the lower portion of the upper- and the mid-tidal flats of these contemporary examples. Mid-flats are characterised by alternating bedload and suspension sedimentation. Sedimentary structures within the sandstones of this facies point to lower flow regime bedload transport with flood and ebb migration of current ripples (Reineck, op. cit.). Cross-bedded units often display complex internal organization with reactivation surfaces (Klein, 1970a, Figs. 27D, 34C) which are attributed to time-velocity asymmetry of tidal currents. Flasers associated with cross-bedding indicate suspension settling of mud during slack water (Reineck and Wunderlich, 1968). Ubiquitous wavy and lenticular bedding are also indicative of alternating bedload and suspension sedimentation. The ripple-cross-laminated sandstone lenses of the lenticular units suggest isolated or "starved" current ripples migrating across a muddy basal layer under the influence of tidal currents (Reineck, op. cit; Klein, 1970b). Interference ripples, in association with the above features, illustrate another diagnostic tidal process, namely ebb-phase runoff accompanying rapid lowering of water-level prior to exposure (Klein, 1963; 1971). The random orientation of current ripple crests (Fig. 52B) provides further evidence of emergence runoff. Desiccation-cracks and intraformational shale-flake conglomerates indicate periodic exposure and evaporation.

The accumulation of muds on tidal flats has been attributed to settling and scour lag effects whereby a suspended particle is transported further shorewards on the slackening current than the position where it first began to settle (i.e. settling lag). The ebb current is then unable to remove all the deposited particles (i.e. scour lag). Moreover, Postma (1967) showed that the time of low current velocity when fine-grained sediments are deposited is considerably longer during high than low tide. McCave (1970) questioned whether accumulations of fine-grained sediments on tidal flats could be deposited from a tidal system, and suggested a periodicity related to storm-calm conditions instead. Storm processes could also be responsible for transporting the sand fraction of lenticular bedded units onto upper tidal flats (Reineck, 1967) while further evidence of short-lived, high-energy processes is provided by the rip-up shale-flake conglomerates of this facies.

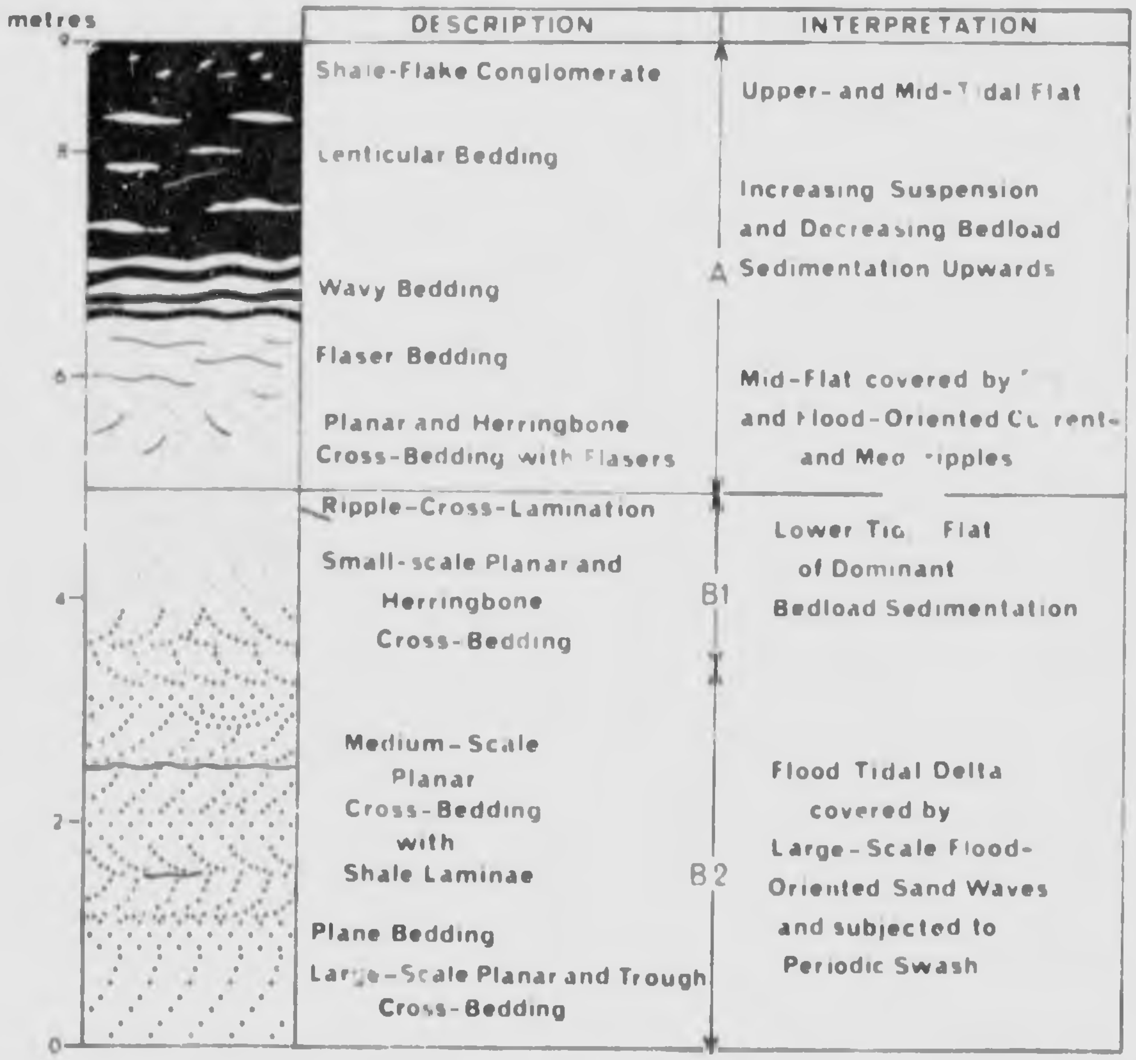

Fig. 60. Palaeoenvironmental interpretation for sandstone-shale (A) and fine- to medium-grained sandstone facies (B1 and B2).

c) Fine- to Medium-grained Sandstone Facies (D2)

The environment of deposition of this facies is interpreted to be basinward of the mid-tidal flat. The abundant small-scale cross-bedding in the first subfacies (Fig. 60, B1) suggests a predominance of lower flow regime bedload processes. Flood and ebb currents appear to have been of equal importance in the migration of current ripples (Fig. 52A). Bedforms of this type are well-documented from lower-tidal sand flats (Reineck, 1967). Thin ripple-cross-laminated units, showing no preferred orientation, often rest on the cross-bedded sandstones. Such cycles of sedimentary structures are similar to the B-C sequences of Klein (1970b), who related interval B to tidal bedload transport and the ripple-cross-laminated unit C to late-stage emergence runoff, with changes in flow direction.

In contrast to current ripples, sand waves and megaripples are not common on tidal flat surfaces, where current velocities are too low for their propagation. High-amplitude bedforms in tidal environments have been observed on tidal deltas (Hayes, et al., 1967; Boothroyd and Hubbard, 1975), in tidal channels (Reineck, 1967) and on shallow marine shelves (Houbolt, 1968). Slip-faces of sand waves on flood tidal delta ramps and channels are oriented landward during both flood and ebb tides (Boothroyd and Hubbard, op. cit.). Flood-oriented cross-beds are produced similar to the large-scale sedimentary structures of this subfacies (Figs. 58; 60, B2). Associated plane-bedding probably formed as a result of high-tide wave swash on the flood ramp. In those portions of flood tidal deltas not shielded from ebb flow, cuspate megaripples display a high bimodality (Boothroyd and Hubbard, op. cit.). The bipolar rose diagram pattern can be considered to reflect such environments. Shale- and siltstone-partings associated with medium-scale trough cross-beds of this subfacies indicate suspension settling in protected portions of flood tidal deltas. The plane-bedded sandstones intercalated within tidal flat sediments of this facies assemblage are interpreted as washover fan deposits. These are related to storm flooding over barrier islands (Hayes, 1967) which results in upper flow regime sheetwash and the generation of plane-bedded sandstones usually with a gentle onshore dip (Andrews and van der Lingen, 1969). The toes of washover fans are commonly structured by landward-dipping planar cross-beds (Schwartz, 1975) but the strike section of available outcrops preclude their exposure in this facies. The underlying multiple-channeled sandstones are more difficult to assign to a specific environment, but probably originated in back-barrier tidal channels, or represent the infilling of scour channels developed during the early stages of washover (Pierce, 1970).

d) Discussion.

The thickness of upward-fining tidal sequences has been used to estimate palaeotidal ranges (Klein, 1972). Based on this reasoning, a macrotidal situation is indicated for this facies assemblage (Fig. 60). Barrier island sequences which are closely-associated with the tidal deposits do not, however, develop under marcotidal conditions (Hayes, 1975) and have already been used to imply a microtidal setting. The presence of washover fan deposits and the lack of back-barrier tidal channel and associated point bar accumulations in this facies assemblage further support a microtidal environment (Hayes, op. cit.). In addition, the limited occurrence of large-scale cross-beds and associated sedimentary structures is compatible with small flood tidal deltas which also characterise microtidal coasts. The exaggerated tidal cycle thickness illustrated in Fig. 60 can thus be interpreted as resulting from stacking of facies.

SPATIAL RELATIONSHIPS OF THE DEPOSITIONAL ENVIRONMENTS.

i) The Steady State Palaeogeographic Model.

Any palaeogeographic model which relates depositional environments in space must be constructed relative to a time plane. Such a datum exists in the Eureka, Saddleback and Stolzburg synclines as the laterally persistent lava at the base of unit MD4 (Fig. 2). Offshore banded iron formations and shales in the Eureka and Stolzburg synclines, and distal alluvial plain siltstones and shales in the Saddleback Syncline, rest on this time plane and provide three reference points for the construction of a palaeogeographic map. The preparation of such a map relies heavily on the interpretation of vertically-stacked sedimentary facies in terms of their lateral distribution at any instant in time. Thus, more proximal barrier island-shallow shelf, as well as deltaic sediments, must have existed on the time plane (palaeoslope) as facies equivalents of distal alluvial plain siltstones and shales in the Saddleback Syncline and of offshore banded iron formations and shales in the Eureka and Stolzburg synclines. Although no volcanic datum exists south of the Saddleback Syncline, the general proximal to distal arrangement of sedimentary facies from south to north suggests that upper- or mid-alluvial plain sedimentation would have occurred at the same time in the Havelock area. The inferred spatial relationships of the different depositional environments in the Moodies Group are illustrated in Fig. 61.*

*FOOTNOTE 2. See Appendix III for palinspastic procedures employed in the construction of the palaeogeographic map.

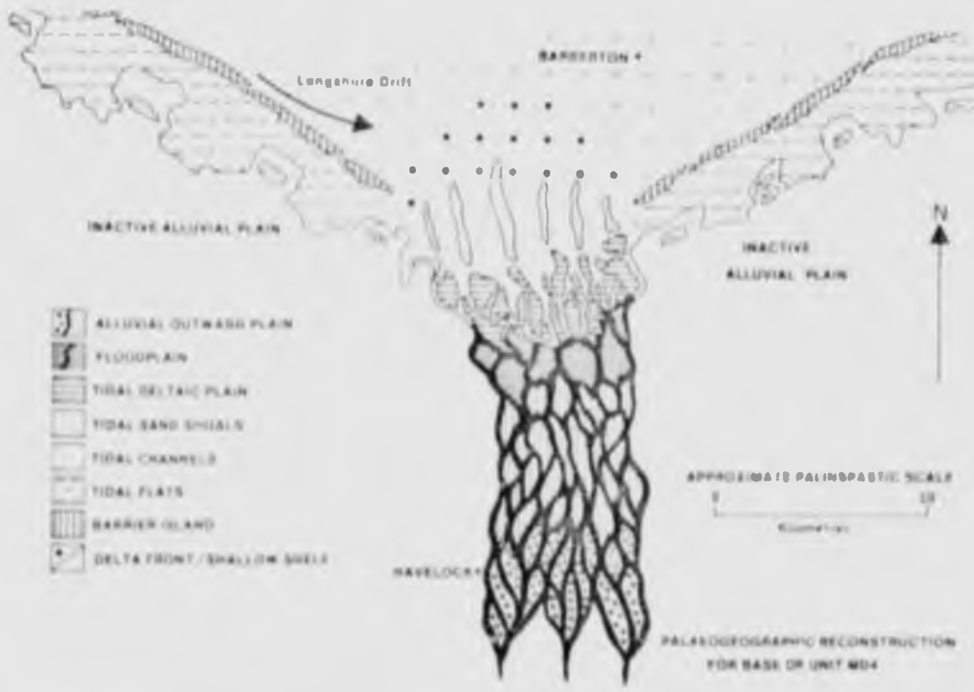

Fig. 61. Inferred palaeogeographic map for base of unit MD4, Moodies Group, Barberton Mountain Land.

The model recognizes the interaction of the four major depositional environments previously proposed, and consists of constructive fluvial and deltaic, and distributive barrier beach-shallow shelf and tidal elements. Upper outwash plain mass flow and conglomeratic channel bar deposits represent the highest energy and most proximal depositional facies. Active vertical accretion of mid-alluvial plain longitudinal bars, accompanied by lateral channel migration resulted in the development of substantial sandstone accumulations. The lower alluvial plain was occupied by high sinuosity streams which developed laterally persistent arenaceous point bar deposits within overbank siltstones and shales. A deltaic plain was influenced by high tidal energies which resulted in the development of tidal channel and sand shoal deposits under macrotidal conditions. Further seawards rapid gravity flow sedimentation occurred in the delta front depositional environment.

Easterly-directed longshore currents redistributed sediments from deltaic to adjacent interdeltaic settings. Elongate barrier islands developed under the prevailing microtidal conditions, with the shoreline profile a response to interacting onshore shoaling waves, longshore currents, offshore rip currents and swash-backwash processes. Easterly-directed longshore currents resulted in the development of laterally persistent, stacked, upward-fining tidal inlet cycles. Ebb and flood currents within tidal inlets had a modifying effect on the shoreface. Low-energy bedload and suspension (with associated chemical) sedimentation characterised the nearshore and offshore shelves respectively. Back-barrier depositional environments were influenced largely by tidal processes, which reworked sediment introduced both from seaward through tidal inlets, and from landward, by weak fluvial processes. Bedload sedimentation predominated on flood tidal deltas and bedload and suspension sedimentation prevailed on tidal flats. Rare washover fan deposits were developed during storms.

The suggestion by Kumar and Sanders (1976) that ancient shoreface sediments are storm phenomena is supported by this investigation. Rip currents, probably generated during storm periods, were the dominant depositional agent in the shoreface. Washover fan sequences in the Moodies Group provide further evidence of storms (Hayes, 1967), while the formation of tidal inlets has also been related to barrier overwash (Pierce, 1970).

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The palaeogeographic model (Fig. 61) suggests the contemporaneous existence of macro- and microtidal conditions along the coastline during the deposition of the Moodies Group. Variations in tidal range along Holocene coastlines (Davies, 1964) can be shown to result from one or a combination of the following factors :

1) Continental shelf width has a significant effect on tidal range; coastlines fronted by narrow shelves have low tidal ranges and those fronted by wide shelves characteristically have amplified tidal ranges. This relationship has been especially well-illustrated along the east coast of the U.S.A. (Redfield, 1958) and also exists off the northwest coast of Australia and the southeast coast of South America.

2) Embayments in the coastline may also modify the tidal range. An important parameter determining the effect of embayments is the ratio $R = L/\lambda$; where L is the distance between the reflecting barrier within the embayment and the mouth of the embayment and λ is the wavelength of the Kelvin or tidal wave rotating about an amphidromic point (Komar, 1976, p. 136). When this ratio is small, tidal ranges within the embayments tend to be the same as those outside. However, if L is larger, and of the order of one-fourth of the tidal wavelength, the embayment resonates with the outside tides, and tidal ranges are up to four times higher, and currents are considerably stronger (Mofjeld, 1976). Holocene embayments such as the Elbe estuary off the coast of Germany, the Gulf of California, the Bay of Bengal, the Gulf of Siam and certain bays in

Australia are characterised by exaggerated tidal ranges without any apparent shelf width influence. The Bay of Fundy, with tidal ranges up to five times those in the adjacent Atlantic Ocean, illustrates this resonant effect particularly well. Coasts off British Columbia and north-east Brazil also have amplified tidal ranges, apparently due to the mildly arcuate concave seaward nature of the coastlines, as the contintental shelves at these localities are exceptionally narrow.

3) Where large quantities of water are introduced into open-ended constrictions, amplified tidal ranges also result. This effect is illustrated in the North Sea, where the east coast of Britain is influenced by macrotidal ranges. Off the coast of east Africa the presence of Madagascar has the effect of generating macrotidal conditions in the Mocambique Channel, while much of the African coast is mesotidal in character.

An embayment model, based on the contemporary German-Danish coastline, and possibly enhanced by a locally wider shelf will account for the coexisting macro- and microtidal conditions inferred in the Moodies palaeogeographic reconstruction (Fig. 61). The macrotidal Elbe estuary contains proximal tidal flats incised by tidal channels and distal sand shoals oriented in an onshore-offshore direction (Reineck and Singh, 1973, p. 318). Barrier islands and back-barrier tidal flats become progressively more prolific to the west and north of the Elbe embayment as tidal ranges decrease and wave processes become increasingly important (Hayes, 1975). The amplification effect of an embayed coastline is also illustrated on the Norfolk and Lincoln coast of England, which serves as a second comparative Holocene model. Except at neap tide, The Wash experiences tidal ranges greater than 5 m (Evans, 1975), while mesotidal conditions exist off northeast Norfolk and result in the development of barrier islands, washover fans and spits (M.R. Leeder, personal communication, 1977).

ii) Stratigraphic Development of the Moodies Group.

a) Introduction.

Stratigraphic sequences in the Moodies Group (Fig. 2) developed as a result of progradation or transgression of the steady state palaeogeographic model (Fig. 61). The intensity of terrigenous input (as controlled by source area tectonics) and rate of basin subsidence determined whether the facies belts migrated landwards or basinwards. Theoretical considerations suggest that :

1) with slow or no subsidence and moderate to rapid sediment influx, progradation of the facies belts would occur;

2) with no subsidence and a slow-to-moderate rate of sediment supply, alluvial facies belts would migrate sourcewards. Similarly, in marine depositional environments, an accompanying gradual relative rise in sea-level, because of sediment compaction, would result in slow transgression; and,

3) with no subsidence and slow sediment influx, a relative rise in sea-level would occur because of sediment compaction and result in rapid transgression (Matthews, 1974, p. 71-72).

b) Progradational Relationships.

Evidence for active progradation, as controlled exclusively by source area tectonics, is illustrated by a number of stratigraphic relationships within alluvial sediments of the Moodies Group. The upper part of unit MD3 in the Saddleback Syncline developed as a result of the northward progradation of proximal and mid-alluvial plain facies across tidal deltaic plain deposits (Fig. 62). Unit MD4 in the same syncline displays evidence of two periods of progradational sedimentation. The lower of the two upward-coarsening sequences consists of distal alluvial plain siltstones and shales, scoured into by proximal channel bar deposits (Fig. 62).

Marine sedimentary sequences in the Moodies Group including deltaic, barrier island-shallow shelf and tidal flat, developed almost exclusively during periods of slow progradation which were separated by rapid transgressive events. The fundamental controls on these contrasting sedimentation styles in each of the marine depositional environments are now considered.

Deltaic deposits in the Moodies Group (lower unit MD3 - Saddleback Syncline) dev[illegible] by seaward progradation despite the intense tidal r[illegible]ng along the embayed coastline. The delta front and [illegible]ging coarse fluvial deposits both provide evidence of an active depositional system during the development of these sediments. This is in accord with modern tide-dominated deltas such as the Ord, Burdekin and Klang, which are all aggrading seawards as a result of moderately high sediment inputs (Coleman, 1976).

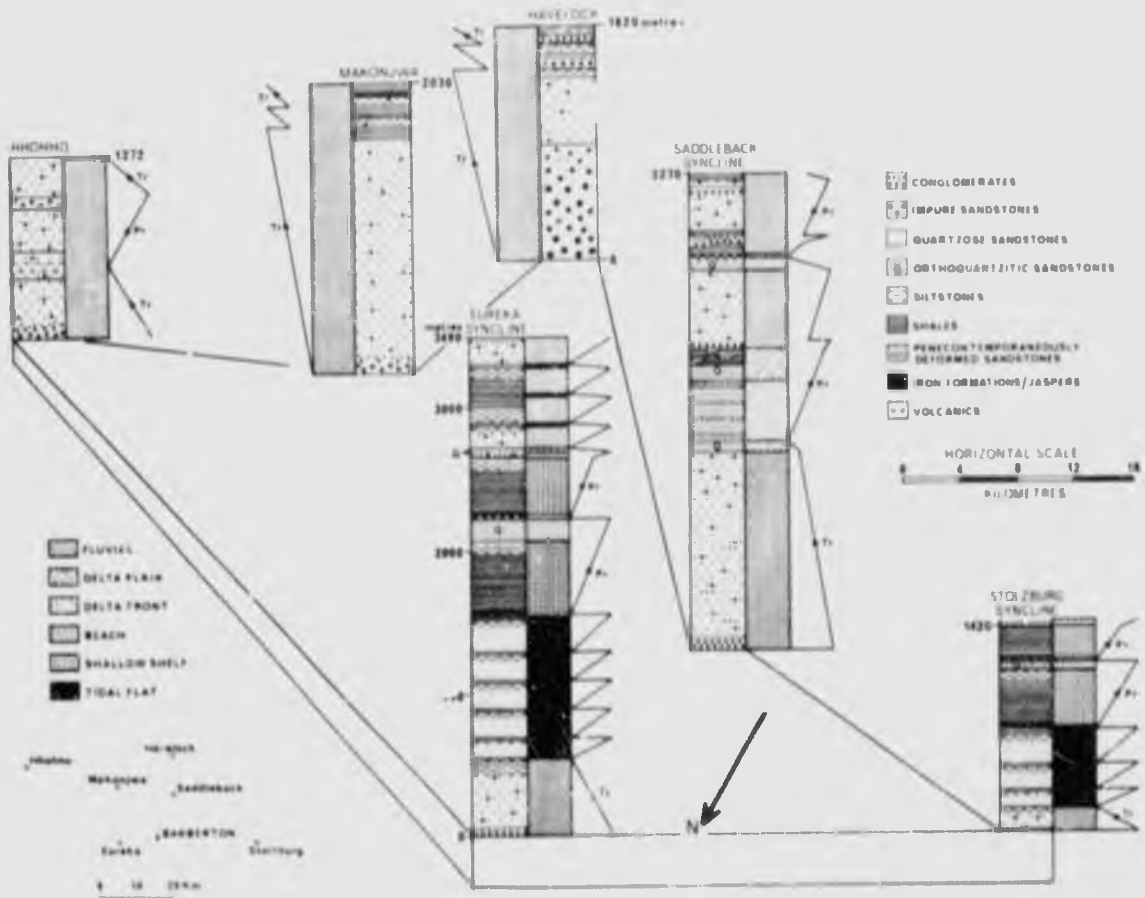

Fig. 62, Depositional environments and progradational-transgressive stratigraphic relationships in the Moodies Group.

The progradational or transgressive nature of barrier islands is determined largely by a surplus or deficit of sand in the shoreface. Under fair weather conditions, shoreface aggradation occurs, while storm processes have the effect of eroding the shoreface. The absence of large-scale landward-dipping longshore bar cross-bedding, indicative of fair weather sedimentation (Swift, 1976), suggests that the rip current deposits in the shoreface sequences of the Moodies Group (Fig. 50) are storm phenomena. Storm-dominated shorelines are thus transgressive unless sand imports excede offshore sand exports (Swift, op. cit.). Sand is supplied to the shoreface by onshore shoaling waves (Clifton et al., 1971), longshore currents (Komar, 1976, p. 217) and jet flow flushing out of sand from estuaries onto ebb tidal deltas (Oertel, 1972). Onshore sand transport by shoaling waves, as well as sand accumulation on ebb tidal deltas, are thought to have been of limited significance along the Moodies coastline. This suggests that longshore sand transport from upcurrent deltas was the dominant import agent, and resulted in a net surplus of sand in the shoreface, which promoted progradation. Seaward accretion rates of 1 km/1000 years are feasible under conditions of moderate sediment supply as documented from Galveston Island by Bernard et al. (1962).

Thicknesses of upward-coarsening deltaic and barrier island deposits have been used to estimate water depths (Klein, 1974). Based on this reasoning, depths of between 400 and 600 m are indicated for the Moodies basin. All depositional facies in these two assemblages are, however, considerab.y thicker than Holocene analogues, and indicate that stacking of facies in a subsiding basin was an important factor. Strong evidence of stacking exists in the case of the spit deposits (Fig. 50). Kumar and Sanders (1974) have shown that spit platform and beachface sequences accumulate between 4 m below mean sea-level and high-tide level. If no stacking occurred in these ancient spit deposits, the 40 m thickness of the plane-bedded orthoquartzitic sandstone facies would imply macrotidal conditions, which are unsuitable for the development of beaches (Hayes, 1975). Thick shoreface accumulations in the ancient have been attributed by Harms (1975) to the preferential preservation of the unusual event. He argued that storm rip currents transport sand further offshore than is the case under normal conditions. Evidence for such high energy processes, during the deposition of the Moodies shoreface sediments has been presented, but their exceptional thickness, coupled with the presence of multiple overlapping tidal inlet cycles, must still be interpreted in terms of facies stacking. Even more convincing evidence of stacking is to be found in the tidal channel deposits, where varied grain-sizes allow for the easy identification of upward-fining cycles. Three or more cycles, each with a basal conglomerate, are often superimposed. The hign degree of maturity of the barrier island and shoal sands, as well as the vertical facies stacking, indicates that basin subsidence closely matched the depositional rate.

Regressive or progradational sequences have been described from a number of Recent tidal flats (Evans, 1975; Reineck, 1975; Thompson, 1975; among others) and attributed to high rates of sediment supply from onshore and/or offshore. Similar conditions must have existed during the development of the upward-fining tidal flat sequences developed in stratigraphic unit MD2 (Fig. 2), and, as has previously been suggested, sediment was introduced onto the tidal flats through inlets, by barrier washover and from the land. Upward-fining cycles of the type shown in Fig. 60 have often been taken as a true indication of tidal range. The possibility of stacking of facies, with resultant exaggerated tidal ranges, has seldom been considered. The recognition of stacking within sedimentary cycles is difficult, but in a tidal environment the composition of the sediments may provide a clue. Tidal sands in non-barred situations are characteristically mature, while those developed behind barrier islands tend to be more argillaceous (Klein, 1967b). All sands of the sandstone-shale and fine- to medium-grained sandstone facies are argillaceous, suggesting limited reworking. The back-barrier depositional setting previously proposed for these sands is thus supported, limiting the tidal range to less than 5 m, and suggesting that the 9 m tidal cycles (Fig. 60) are exaggerated and probably a result of stacking.

Under conditions of maximum sediment input, mid-alluvial plain conglomerates and sandstones prograded directly onto nearshore siltstones and sandstones (Fig. 63) in the form of fan deltas (Flores, 1975; Galloway, 1976). This is well-illustrated at the top of unit MD4 in the Eureka and Stolzburg synclines (Fig. 2). Back-barrier tidal flat and barrier island depositional environments were not developed under these conditions of active sediment supply.

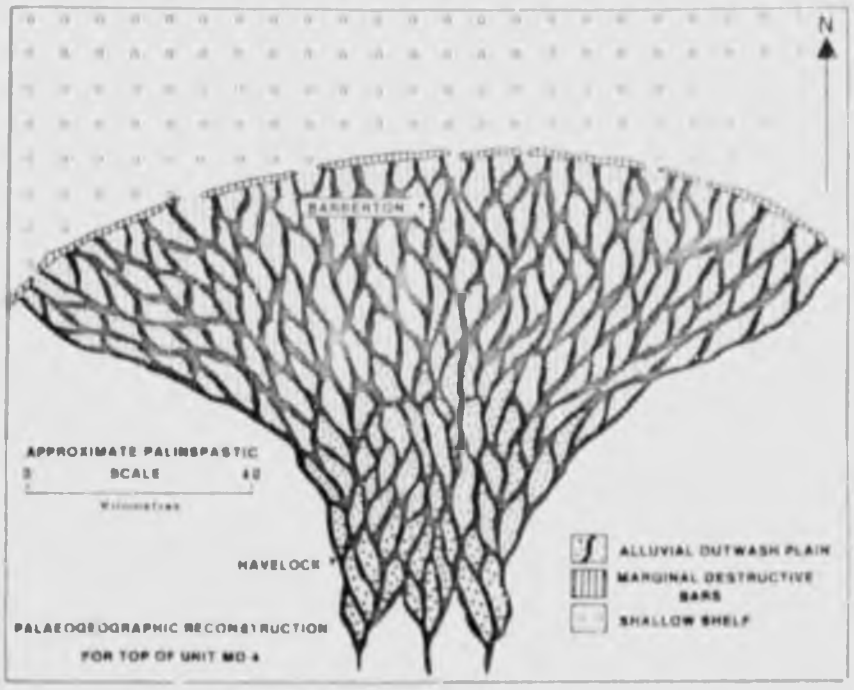

Fig. 63. Palaeogeographic fan delta model for an actively prograding alluvial sy   upper Unit MD4; Moodies Group, Barberton Mountain Land.

c) Transgressive Relationships.

Stratigraphic unit MD1, throughout the Barberton Mountain Land, is thought to have developed as a result of the slow sourceward retreat of alluvial plain facies belts. This was in response to decreasing uplift i the source area, although pebble-types in the basal conglomerate, such as the granitic clasts in the Eureka Syncline, often reflect derivation from local floor irregularities.

Transgressive stratigraphic relationships in the Moodies Group most commonly reflect sudden, rather than gradational changes of depositional environment. These can be ascribed to a cessation of sediment supply in response to decreasing tectonic activity in the source area or, as appears more probable, a change in locus of deposition (Asquith, 1974). Close examination of Figs. 2 and 62 together reveals the following transgressive relationships :

1) Stratigraphic unit MD2 in the Eureka and Stolzburg synclines consists of stacked upward-fining tidal flat cycles (see Fig. 60) periodically interrupted by washover fan sequences. Periods of slow progradation separated by rapid transgressive events are indicated.

2) Tidal flat sequences are directly overlain by shallow shelf deposits at the top of unit MD2 in the Eureka and Stolzburg synclines. The absence of barrier sediments along this transgressive surface suggests rapid drowning as a result of sudden deepening of the water-level (Sanders and Kumar, 1974) and, as shown by Swift (1968) and Kraft (1971), is not unexpected as transgressive barrier island deposits have extremely low preservation potentials. Along storm-influenced coastlines in particular, barrier sediments are reworked and incorporated into storm washover and tidal flat deposits.

3) Barrier spit sequences are directly overlain by shallow shelf deposits at the top of unit MD3 in the Eureka and Stolzburg synclines. No transgressive basal conglomeratic lag deposits, of the type reported from Tertiary barrier island sequences in California (Clifton, 1973a), occur, suggesting very rapid transgression, possibly associated with sudden 'jumps' of the shoreline (Sanders and Kumar, 1974).

The rise in sea-level in each of these three cases was probably relative, as opposed to absolute, inasmuch as sediment compaction possibly associated with basin subsidence, during periods of no fluvial, onshore or longshore sediment supply would have promoted transgression.

4) Plane-bedded orthoquartzitic sandstones, already interpreted as swash deposits, overlie 'impure' fluvial sandstones at the top of units MD1 and MD4 in the Saddleback Syncline. The same stratigraphic intervals in the Eureka and Stolzburg synclines are planes of rapid transgression (Fig. 62). A 10 to 20 km landward 'jump' of the breaker zone is thus indicated with the plane-bedded orthoquartzitic sandstones representing transgressive sheet sandstones developed by reworking of the underlying fluvial deposits along the advancing shoreline. The poorly-sorted fluvial and mature beach (swash zone) sandstones at the top of unit MD3 in the Saddleback Syncline are clearly separated on a plot of mean grain diameter against grain-size standard deviation for samples from the two inferred environments (Fig. 64). The samples lie within or close to their respective fields as delineated by Molola and Weiser (1968) using samples from known Holocene depositional environments. Marine reworking of fluvial pebbles and boulders is thought to have led to the development of the vertically discrete conglomerates intercalated within clean sandstones, at the top of unit MD3 in the Saddleback Syncline. Similar onlap relationships have been documented from the Mississippi Delta. Following abandonment of individual lobes and compaction of the sediment pile, clean sandstones are developed

during transgressive wave reworking of the deltaic plain (Scruton, 1960; Kolb and Van Lopik, 1966; Morgan, 1970). McGowen (1970) and Stephens et al. (1976) have observed identical destructive relationships for the Gum Hollow fan delta and Santee River delta, respectively.

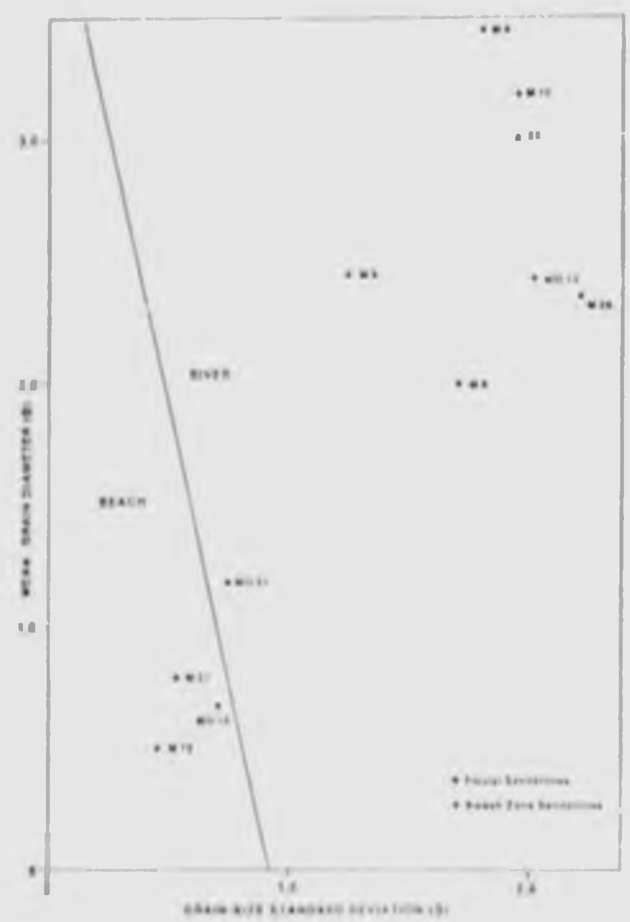

Fig. 64. Plot of mean grain diameter against grain-size standard deviation for 'impure' fluvial and beach (swash zone) sandstones. Boundary between fluvial and beach fields after Moiola and Weiser (1968). Sample localities are described in Tables 3 and 4.

5) Much of unit MD5 in the Eureka Syncline consists of interlayered siltstones and shales with minor sandstones (Fig. 2) which have already been interpreted as distal alluvial plain channel-fill and overbank deposits (Fig. 20). A number of intercalated quartzose sandstones contain sedimentary structures such as herringbone cross-

bedding and abundant shales flasers (Fig. 65) which provide evidence of transgressive reworking of alluvial plain sediments by tidal processes. In many respects this is an analogous relationship to that described by McGowen (1970) from the Gum Hollow fan delta, where distal surficial fan sediments are being reworked in the intertidal zone to form a thin veneer of fine-grained wave-rippled and plane-bedded sands.

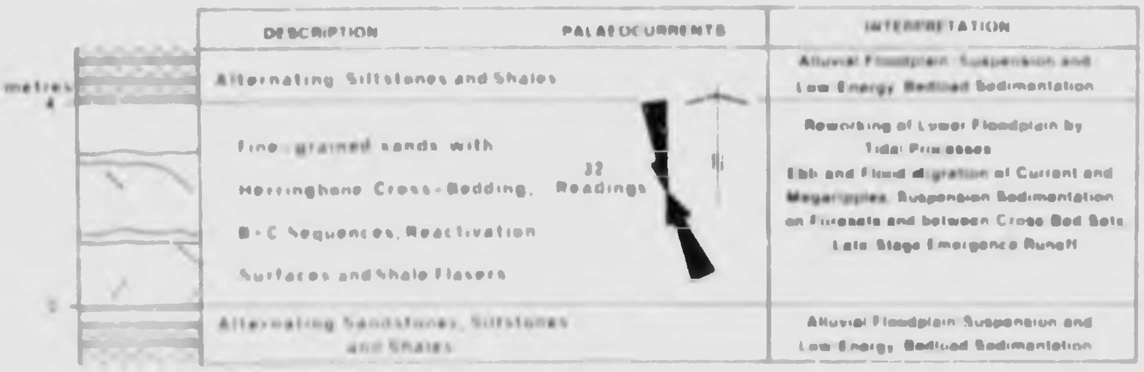

Fig. 66. Idealized stratigraphic column showing inferred tidal flat reworking of distal alluvial plain sediments.

d) <u>Discussion</u>.

From the foregoing discussion it can be appreciated that the preservation of different depositional facies in the steady state model (Fig. 61) was determined to a large extent by the transgressive or regressive nature of the shoreline. Under transgressive conditions, barrier island deposits were mostly destroyed, and less commonly reworked into thin orthoquartzitic sandstones (Fig. 62). Complete upward-coarsening barrier island and deltaic sequences require a regressive shoreline for their development and preservation. Abundant sediment supply was invoked as a prerequisite for progradational sedimentation along a regressive shoreline. Tidal flat deposits had their highest preservation potential in a general transgressive situation, although individual upward-fining cycles developed under progradational conditions.

iii) The Fig Tree Enigma.

There can be no doubt that the Fig Tree sediments, as preserved today (Fig. 1) are older than the Moodies. What is less clear, however, is whether these two sequences belong to a single depositional cycle, or to pre- and post-orgenic flysch-and molasse-type cycles, respectively, as first suggested by Anhaeusser et al. (1968). Before this problem can be resolved, the character of the Fig Tree sediments, as well as the nature of the contact between the Moodies and Fig Tree groups, must be considered.

Greywackes are the predominant lithology in the Fig Tree Group, and comprise greater than 80 per cent of the lower half of this unit. Quartz, potash-felspar and chert are the predominant framework constituents, while subsidiary lithic particles are of volcanic, sedimentary, plutonic and metamorphic derivation. The greywackes contain in excess of 30 per cent clay-matrix, and up to 8 per cent diagenetic dolomite (Reimer, 1975). Shale, chert and banded iron formation become more abundant upwards, along with a sympathetic decrease in grain-size of the greywackes (Van Vuuren, 1964). Local small-pebble and resedimented greywacke conglomerates occur throughout the group, and lens out along strike (Anhaeusser, 1976). Greywacke grits at the top of the succession characteristically contain well-rounded 1 - 5 mm quartz grains which contrasts with the angular nature of framework constituents in the underlying greywackes. Geochemical studies (Condie et al., 1970) have revealed that granitic components increase upwards in the Fig Tree Group at the expense of more mafic and ultramafic elements.

Sediments of the Fig Tree Group are ubiquitously graded, in units from a few inches to several feet in thickness (Van Vuuren, 1964). Complete Bouma (1962) cycles are rarely observed; instead a coarse, graded basal unit is overlain by plane-bedded fine-grained sandstones and siltstones, followed by an argillaceous layer. The most complete cycles are capped by cherts or banded iron formations (Reimer, 1975). Frequently, however, the grading is only apparent through the presence of thin argillaceous layers on top of the greywacke units (Van Vuuren, op. cit.). Sole markings, flute casts, load casts and flame structures are frequently developed at the base of graded units where in contact with shale (Anhaeusser, 1976). Small-scale slump structures are often present within the greywackes. In the absence of cross-bedding, palaeocurrent determinations have been based on the orientation of long axes of flute casts and sole markings. These have revealed that dispersal currents operated in a general northwesterly to northeasterly direction (Reimer, op. cit.).

Sediments of the Fig Tree Group can be accommodated within Turner and Walker's (1973) 'resedimented association'. The greywackes and conglomerates were formed by turbidity current resedimentation from shallow- into deep-water, and the shales, cherts and banded iron formations developed as a result of background, deep-water suspension sedimentation and chemical depositional processes. Turbidite sediments owe their origin to subaqueous, deep-water gravity flow processes, and turbidity currents most commonly develop in submarine canyons (Klein, 1975). Sufficiently detailed studies to distinguish between regional slope, turbidity feeder channel and proximal and distal submarine fan facies have not yet been undertaken on the Fig Tree Group, but it is reasonably safe to conclude that this sequence accumulated in neritic and/or bathyal environments. The lower Fig Tree greywackes, shales and cherts may represent the most distal sediments passing upwards into regional slope, predominantly argillaceous and chemical deposits, with the greywacke grits at the top of the succession providing evidence of shallow-water reworking immediately prior to the deposition of the Moodies Group (Van Vuuren, 1964). A similar turbidite to shallow-water transition is described from the Namurian Mann Tor Sandstone and Shale Grit turbidite of northern England (Walker, 1969).

Contrary to the often indirect evidence proposed by Visser (1956), for an unconformity between the Moodies and Fig Tree groups this contact is most commonly transitional. This is best seen along the northern and southern limbs of the Eureka Syncline while detailed mapping in the Saddleback and Makonjwa areas (Bell, 1967; Fig. 1) has revealed a similar relationship. Available field evidence thus suggests that the Moodies and Fig Tree groups were formed during a single major depositional cycle, and it is suggested that the latter simply represents a deeper-water facies of the continental and shallow-marine Moodies Group.

Active progradation towards the north was responsible for the observed stratigraphic relationships between the two groups. Additional support for this proposal comes from a common source area to the south for these two sedimentary sequences (Reimer, op. cit.). Furthermore, the Fig Tree greywackes are lithologically very similar to the little-reworked fluvial sediments in the Moodies Group.

DISCUSSION.

i) General.

The present study has answered a number of important questions pertaining to Archaean sedimentation in the Barberton Mountain Land, South Africa. These have enabled considerable refining of the 'continental association' (Turner and Walker, 1973) to be achieved and include :

a) generalizations relating to Archaean sedimentary facies-types in the Barberton Mountain Land, and grouping of these into widely recognizable associations;

b) an understanding of the processes responsible for the deposition of clastic Archaean sediments in the Barberton Mountain Land; and,

c) an understanding of the evolution of the Archaean sedimentary basin in the Barberton Mountain Land, through examination and interpretation of vertical facies sequences and lateral facies changes.

ii) Basin Geometry.

An appreciation of the geometry of the depositional basin is essential and is one broader implication of the study which is developed below :

a) Physical sedimentary processes in Archaean marine environments were remarkably similar to those operating along present-day coastlines. Continental sediments were deposited in environments very similar to contemporary non-vegetated outwash fans, such as those which occur in Alaska.

b) Facies analysis of clastic sediments in the Swaziland Supergroup has revealed a smooth, continental to marginal marine to deep-water marine transition. The suggestion by Pettijohn (1972) that Archaean depositories had elongate, possibly fault-bounded, trough-like geometries does not appear to apply to the Moodies Group. Rather, the Moodies sediments were derived exclusively from the south and deposited, in part, on a stable continental shelf.

The above criteria can be used to suggest that the Moodies and Fig Tree sediments accumulated along an ancient continental margin deepening to the north. Highly tentative proposals along parallel lines were previously made by Goodwin (1974) and Windley and Smith (1976) for sediments of Archaean age The present detailed facies analysis has revealed many similarities between Archaean sediments in the Barberton Mountain Land and deposits along Holocene continental margins. Voluminous sediment supply results in the outward construction of a progradational continental embankment composed of Moodies-type continental and marginal marine sequences (Dickinson, 1974). For example, the outer-edge of the embankment off the Niger Delta contains depositional phases analogous to those previously described from the Fig Tree Group, namely, a basal phase of sandy turbidites near the toe of the embankment, a middle phase of largely argillaceous sediments deposited on the advancing frontal slope of the embankment, and an upper phase of largely arenaceous shallow marine strata deposited along the outer edge of the top of the embankment (Burke, 1972).

Any hypothesis regarding the exact nature of the continental margin should be developed within the context of existing models for greenstone evolution. One of these invokes rifting of an older sialic crust, and the accumulation of volcanics and sediments in grabens (see example Hunter, 1974 ; Archibald et al, in press). While the latter may well account for the lower volcanic sequences in Archaean greenstone belts, the sedimentary facies patterns recognized for the Moodies and Fig Tree Groups cannot be reconciled with an elongate, fault-bounded, trough-like depository. Continued rifting, followed by separation of sialic blocks will, however, result in the development of a stable continental coastal margin (McCrossan and Porter, 1973) analogous to that envisaged as the depositional setting for these Archaean sediments.

A second model draws on a complex of island arc tectono-volcanic and sedimentary processes (see for example Engel and Kelm, 1972; Anhaeusser, 1973) and suggests that sedimentation would have occurred in interarc or backarc basins, similar to those of the western Pacific (Goodwin, 1974). Although sedimentation trends exhibited by the Moodies and Fig Tree groups are compatible with a backarc depositional setting, this model for greenstone belt evolution has recently been criticized by Burke et al. (1976) and Tarney et al. (1976) on the grounds that the petrology and chemistry of the volcanic rocks display affinities not only with lavas extruded along island arcs above subduction zones, but also with eruptives at mid-oceanic ridges. These same authors, along with Windley and Smith (1976), have suggested instead that Archaean greenstone belts are comparable to marginal basins flanked on one side by a stable, rifted continental margin (possibly analogous to that inferred to have existed to the south of the Barberton Mountain Land) and along which sediments of the Moodies and Fig Tree-types may have accumulated. The opposing island arc flank would have been characterized by volcanogenic sedimentation and, if represented in the Archaean of South Africa, will exist in greenstone belts north of the Barberton Mountain Land. Further research along these lines is anticipated.

Variations in tidal range along the Moodies palaeocoastline are reconcilable with the suggested continental margin depositional environment. The widespread evidence of microtidal conditions in the Moodies Group, coupled with the presence of coeval deep-water turbidite sediments in the Fig Tree Group, further suggest a narrow continental shelf at the time of deposition of these Archaean sediments. Local widening of the shelf and/or the occurrence of embayments along the coastline would explain the development of the localized regions of macrotidal range. Coleman (1976) suggested that ancient, tide-dominated deltas would have occurred along the margins of stable platforms, where tidal ranges would have been significantly amplified.

iii) Comparison with Proterozoic Sedimentation Styles.

The style of sedimentation in certain Proterozoic stratigraphic units on the Kaapvaal Craton, such as the Pongola Supergroup (Von Brunn and Hobday, 1976) and Pretoria Group (Button and Vos, 1977), differs significantly from that in the Archaean. Not only are deep-water sediments absent from these sequences, but widespread evidence of macrotidal conditions also exists. In addition, barrier-beach depositional assemblages are lacking from these Proterozoic sedimentary sequences. Broad, stable intra-cratonic shelves are indicated, a feature which has previously been considered as characteristic of the Proterozoic (see for example Brooks et al., 1976); and contrasts sharply with the narrow continental shelf which existed during the deposition of the Archaean Moodies and Fig Tree groups. The evidence of high tidal ranges in the above Proterozoic sedimentary basins implies that they were connected to an open oceanic body of water. Not all sediments on the Kaapvaal Craton, however, were deposited under macrotidal conditions. Evidence of low tidal ranges exists for the chemical sediments in the lower part of the Transvaal Supergroup (Beukes, 1977) and in the Waterberg Supergroup (Vos and Eriksson, 1977). These sequences, as well as the Upper Witwatersrand alluvial plain-lacustrine deposits (Vos, 1975), probably accumulated either within enclosed intracratonic basins having no connection to the open sea, or in depositories which were too shallow for amplification of the tidal wave to occur.

iv) Broader Implications.

The few detailed sedimentological analyses of Archaean sequences elsewhere in the world, which are lithologically similar to the Moodies and Fig Tree groups of the Barberton Mountain Land, have suggested that turbidite and alluvial deposits predominate (see introduction for specific references). With the exception of Donaldson and

Platt's (1975) study in the Yilgarn Block of Western Australia, no marginal marine sediments of the Moodies-type have been recognised. This may be largely a function of inadequate palaeoenvironmental analyses, although Turner and Walker (1973) have provided convincing evidence that turbidite and alluvial fan sediments alone occur in the Sioux Lookout greenstone belt in Canada. Available information on other sequences suggests that a second style of sedimentation, analogous to that proposed by Pettijohn (1972), may have existed during the Archaean, and involved high energy alluvial and turbidite sediments being fed into localized elongate, trough-like depositories from proximal source areas.

If Archaean sedimentation was associated with rifted continental margin tectonics the two sedimentary styles may reflect :

a) an early response to rifting, as typified by the Archaean sediments of Canada; and,

b) a response to more stable continental margin conditions after rifting and opening of the sea, as represented by the Moodies and Fig Tree Groups.

Archaean sediments in Australia and Rhodesia, in particular, require detailed examination employing modern sedimentological techniques to test this hypothesis.

CONCLUSIONS.

i) This palaeoenvironmental analysis has revealed that physical depositional processes during the Archaean were remarkably similar to those operating today. Evidence of fluvial, tidal, wave swash, rip and long-shore current, storm and turbidity processes have been recognized.

ii) The recognizable similarities between Archaean and Holocene depositional processes has led to the conclusion that the Moodies and Fig Tree groups accumulated along an ancient continental margin.

iii) The widespread evidence of microtidal sedimentation during the deposition of the Moodies Group suggests the existence of a narrow continental shelf. Embayments along the coastline and/or a widening of the continental shelf generated localized macrotidal conditions, under which sand shoals developed perpendicular to the shoreline at the expense of beaches.

iv) The abundant development of tidal sediments in the Moodies Group indicates that the Earth-Moon system was in existence at least 3 000 m.y. ago.

v) The quartzo-felspathic nature of arenaceous sediments in the Moodies Group indicates the widespread existence of granitic rocks at the time of deposition of these Archaean sediments. Unequivocal evidence for the existence of emergent landmasses to the south of the outcrop belt is provided, and contradicts the suggestion of Hargraves (1976) that a primordial sea covered the earth until 2 000 m.y. ago.

vi) In contrast to the red colouration exhibited by fluvial sediments less than 2 200 m.y. in age (see for example Cloud, 1976), those in the Moodies Group are ubiquitously dull in colour, indicating an anoxygenic atmosphere at the time of their deposition. Chemical weathering during the Archaean was achieved principally through carbonation and hydration.

vii) This study has revealed the existence of widespread marginal marine deposits, and contrasts with most other palaeoenvironmental investigations on sediments of this age, which have suggested that alluvial and turbidite sedimentation may have characterised the Archaean. Detailed palaeoenvironmental investigations of sedimentary sequences in Australia and Rhodesia are necessary to determine whether two distinct depositional styles may have existed during the Archaean; namely, open continental margin and elongate, trough-like basin sedimentation. If this can be shown to be the case, it may be possible to relate these two styles of sedimentation to different stages of continental margin rifting.

viii) Detailed palaeoenvironmental analyses of Precambrian sedimentary basins can provide important information pertaining to crustal evolution. Whereas a narrow continental shelf existed during the accumulation of the Moodies and Fig Tree groups, a number of sedimentary sequences on the Kaapvaal Craton contain evidence of widespread macrotidal conditions, suggestive of deposition on a broad, stable shelf.

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APPENDIX I

A. Calculation of Statistical Parameters (Inman, 1952)

i) Mean $= \dfrac{\phi_{16} + \phi_{84}}{2}$

ii) Standard Deviation $= \dfrac{\phi_{84} - \phi_{16}}{2}$

B. Thin Section - Sieve Conversion Equations (Harrell and Eriksson, in preparation)

i) $\phi_{2_S} = .16412 + 1.13674 \times \phi_{2_T}$

$n = 51 \quad r = .983$

ii) $\phi_{5_S} = .15646 + 1.09672 \times \phi_{5_T}$

$n = 66 \quad r = .973$

iii) $\phi_{9_S} = .17751 + 1.06609 \times \phi_{9_T}$

$n = 72 \quad r = .975$

iv) $\phi_{16_S} = .12722 + 1.07477 \times \phi_{16_T}$

$n = 73 \quad r = .977$

v) $\phi_{25_S} = .11671 + 1.06432 \times \phi_{25_T}$

$n = 72 \quad r = .976$

vi) $\phi_{36_S} = .093944 + 1.05452 \times \phi_{36_S}$

$n = 73 \quad r = .965$

vii) $\phi_{50_S} = .12098 + 1.03027 \times \phi_{50_T}$

$n = 75 \quad r = .958$

viii) $\phi_{64_S} = .25454 + .96924 \times \phi_{64_T}$

$n = 71 \quad r = .957$

ix) $\phi_{75_S} = .32539 + .95237 \times \phi_{75_T}$

$n = 68 \quad r = .917$

x) $\phi_{84_S} = .45205 + .89492 \times \phi_{84_T}$

$n = 62 \quad r = .912$

xi) $\phi_{91_S} = .57891 + .84562 \times \phi_{91_T}$

$n = 53 \quad r = .862$

xii) $\phi_{95_S} = .77240 + .80095 \times \phi_{95_T}$

$n = 41 \quad r = .830$

xiii) $\phi_{98_S} = .98885 + .76964 \times \phi_{98_T}$

$n = 23 \quad r = .723$

n = number of paired thin section-sieve samples used in calculation of equations.

r = correlation coefficient.

s = sieve.

t = thin section.

Cumulative percentiles greater than 84 could not be read from the impure sandstone curves with any degree of accuracy (see Fig. 8). Coupled with the poor correlation coefficients for the high percentile coversion equations (see above), this determined that the statistical parameters had to be calculated using the equations of Inman (1952) instead of Folk (1968).

APPENDIX II

Procedure employed for field measurement and stereonet rotation of Palaeoccurent Data.

A. FIELD MEASUREMENT.

i) Planar Cross-Beds :

Measure strike and dip of bedding plane and strike and dip of foreset surface.

Example :-	Bedding	:	$275^{\circ}/80^{\circ}S$
	Cross-Bedding	:	$255^{\circ}/55^{\circ}S$

ii) Trough Cross-Beds and Ripple Marks :

Measure strike and dip of bedding plane and pitch of trough cross-bed axis or ripple crest in the bedding plane. Also record whether trough is directed into or out of the ground.

Example :-	Bedding	:	$275^{\circ}/80^{\circ}S$
	Pitch of Trough Axis	:	$70^{\circ}W$ into the ground.

B. DETERMINATION OF ATTITUDE OF FOLD AXIS OR AXES.

Example :- Eureka Syncline F_1 Axis : 50° Plunge on 255° (in eastern segment of fold); F_2 Axis: 70° Plunge on 160° (Data after Anhaeusser, 1976).

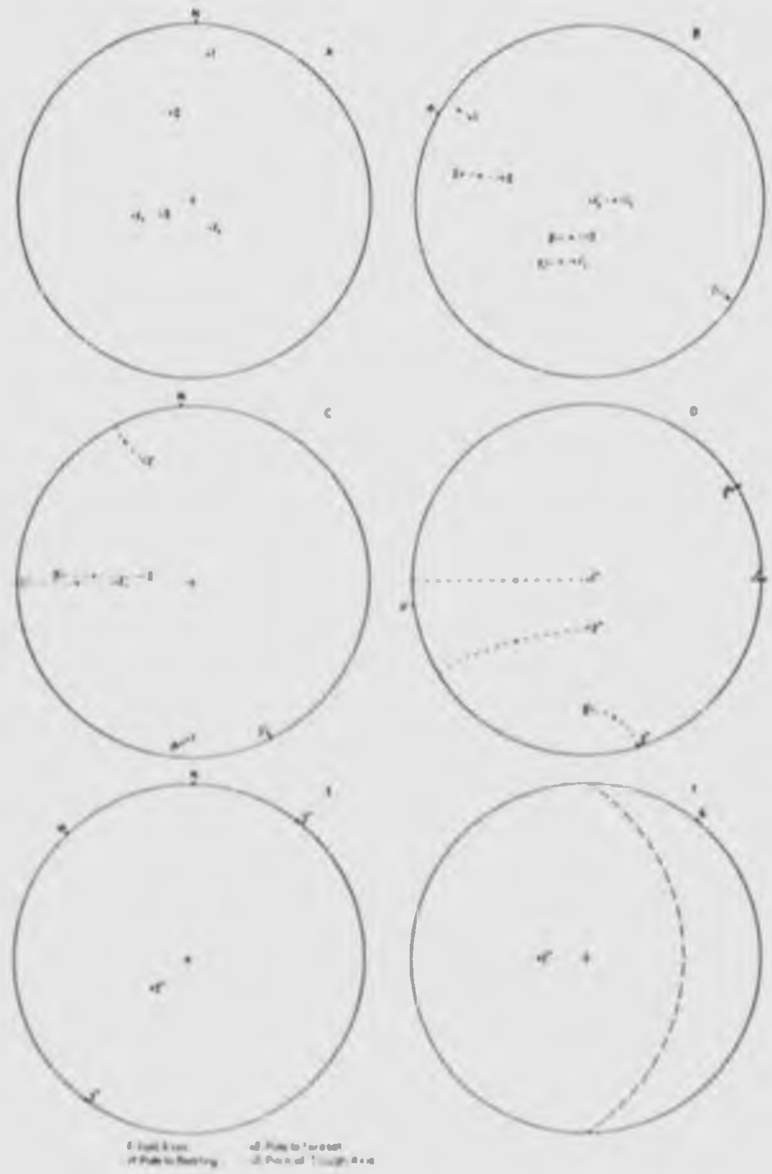

Fig. 66. Stere[illegible] [illegible]ional procedures for palaeocurrent data.

C. PLOTTING OF DATA ON STEREONET.

(Refer to Figure 66A)

1. : Pole to Bedding Plane.

2. : Pole to Foreset Surface.

3. : Pitch of Trough Axis in Bedding Plane.

F_1. : First Fold Axis.

F_2. : Second Fold Axis.

D. ROTATIONAL PROCEDURES.

Step 1 : Refer to Fig. 6´B. Rotate F_2 to vertical and other points by 20^o along small circles to 1', 2', 3', F_1' and F_2'.

Step 2 : Refer to Fig. 66C. Rotate F_1' to horizontal and all other points by 42^o along small circles to 1", 2", 3" and F_1".

Step 3 : Refer to Fig. 66D. Rotate bedding plane to horizontal (i.e. pole to bedding to vertical). Pole to foreset (2") and pitch to trough axis (3") are rotated by equal amounts to 2‴ and 3‴ .

Step 4 : Refer to Fig. 66E. Axial plane trace in eastern segment of Eureka Syncline is oriented approximately east-west as a result of tectonism post-dating F_1. Axial plane traces elsewhere in the Barberton Mountain Land trend approximately northeast-southwest. Stereonet must therefore be rotated by 45^o in an anticlockwise direction to eliminate the effects of buckling associated with F_2. This is achieved by generating a new north direction (N') on the stereonet.

NOTE: Exclude steps 1 and 4 if strata were subjected to only one phase of deformation. This is the case in all other parts of the Barberton Mountain Land.

E. READ DIRECTION OF BEDFORM MIGRATION OFF THE STEREONET.

i) Read trend of trough axis relative to N' off Fig. 66E. Answer: 038^o or 218^o. Trough is directed into the ground. Trough cross-bed was therefore formed by megaripple migration in direction N38^{0}E.

ii) Construct strike and dip of foreset surface using pole 2"'. Read attitude of this surface off Figure 66F.

F. PRESENTATION OF DATA.

Palaeocurrent cross-bed data are presented in rose diagram form.

APPENDIX III

Palinspastic Procedures employed in the construction of the Palaeogeographic Maps. (Figs. 55, 63.)

A. GENERAL.

Strata of the Moodies Group occupy a series of north easterly-trending isoclinal synclines (Fig. 1) which are approximately parallel to the depositional strike. For tightly folded strata of this type, palaeogeographic maps, portraying the spatial relationships of depositional environments, can only be constructed if :

i) limbs of folds are rotated to the horizontal;

ii) some estimation of the distance between unfolded synclines can be made.

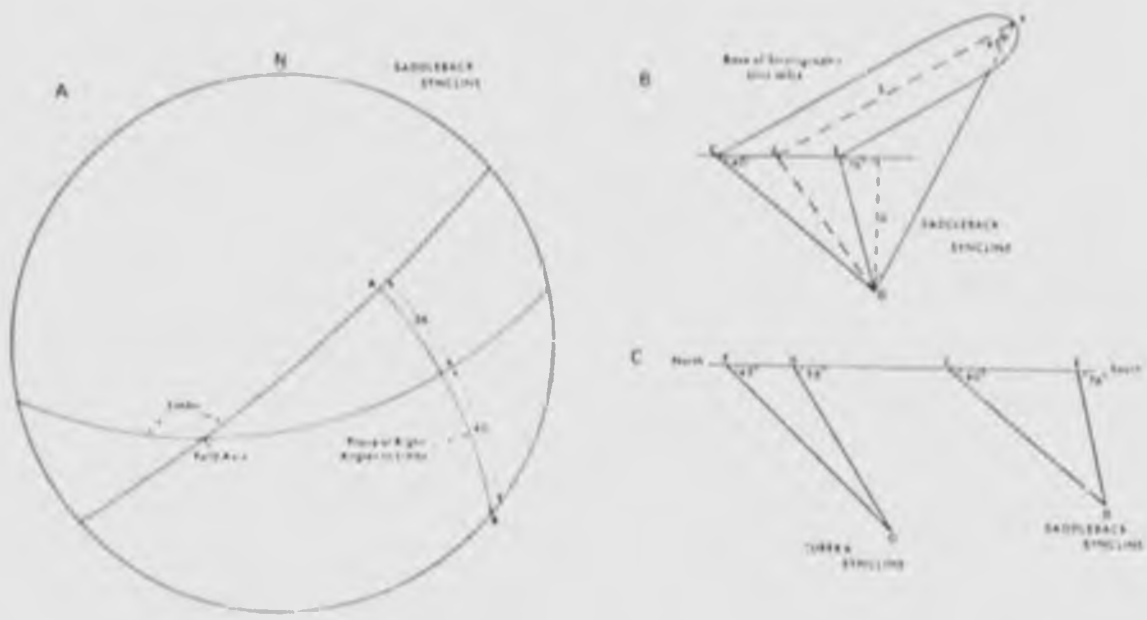

Fig. 67. Stereonet and geometrical procedures for unfolding limbs of folds.

B. ROTATION OF LIMBS TO THE HORIZONTAL.

i) Plot strike and dip of bedding planes of each fold limb.

Example: Saddleback Syncline:- N Limb - $080^\circ/55^\circ S$
S Limb - $050^\circ/80^\circ S$

Intersection of the two planes defines the fold axis (refer to Fig. 67A).

ii) Construct a plane A - B at right angles to the limbs.

iii) Read pitch of intersection of the two limbs with this plane. Pitches are respectively $76^\circ S$ and $40^\circ S$.

iv) Measure distance X - Y (S) from fold closure to position of control locality (X) for any given stratigraphic horizon (refer to Fig. 67B).

Example: Palaeogeographic map (Fig. 61) constructed for base of stratigraphic unit MD4.

v) Calculate

$$d = S \sin 45^\circ = 5.75 \sin 45^\circ = 4.07 \text{ Km}$$

vi) Calculate distances :

$C - D = d \operatorname{Cosec} 40^\circ = 6.33$ Km; and,

$D - E = d \operatorname{Cosec} 76^\circ = 4.19$ Km

The sum of C - D and D - E corresponds to the distance around the limbs of the fold.

vii) The same procedures are carried out for each fold.

C. ESTIMATION OF DISTANCES BETWEEN UNFOLDED SYNCLINES.

Individual synclines of unfolded Moodies sediments are separated by complexly folded Onverwacht and Fig Tree strata, and often by major strike faults. The second stage of the palinspastic procedure is thus imprecise. As a first approximation it has been estimated that the amount of Moodies sediments removed by erosion of intervening anticlines roughly equalled the preserved distance around the limbs of the adjacent synclines.

Example : Palinspastic reconstruction for the base of stratigraphic unit MD4 in the Eureka and Saddleback Synclines (refer to Fig. 67C).

Distal alluvial plain sediments now exposed at the base of stratigraphic unit MD4 on the southern limb of the Saddleback Syncline graded over a distance :

ED + 2(DC) + 2 (GH) + FG

= 2.66 + 2(3.31) + 2(6.33) + 4.19

= 26.13 Kilometres

into offshore banded iron formations and shales at the base of the same stratigraphic on the northern limb of the Eureka Syncline.